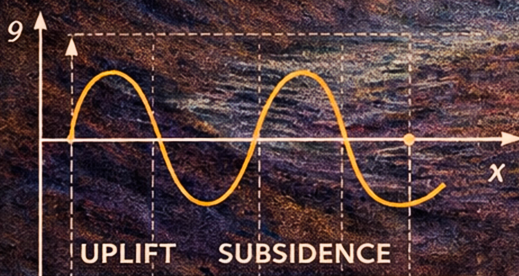


V.V. BELOUSOV

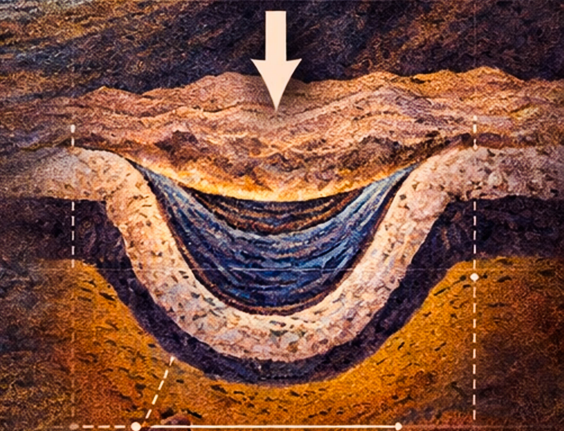
BASIC PROBLEMS IN GEOTECTONICS

A SYNTHESIS OF
VERTICAL CRUSTAL OSCILLATIONS
AND GEOSYNCLINAL THEORY

OSCILLATING CRUST



$$\Sigma = \min \Sigma (y_i - B_{is} - B_{ixx})^2$$



Geosyncline

Paleozoic

Mesozoic

Basic Problems in Geotectonics

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Basic Problems in Geotectonics

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BASIC PROBLEMS IN GEOTECTONICS

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PREFACE

Geotectonics is the branch of geology that deals with the movements and deformations of the earth's crust. Since many other geologic processes, both within the depths of the earth and on its surface, are closely associated with movements of the earth's crust, this branch of geology has a somewhat peculiar status and provides, in some measure, the theoretical basis of geology.

It is the basic task of any scientific discipline to explain the laws governing the occurrence and development of the processes it studies. Likewise, *the most important task of geotectonics is to establish the laws that govern the development of movements and deformations of the earth's crust*. It is to this task that the present book has been devoted.

The author has tried to achieve the following:

To show that geotectonics is an inductive science based on data obtained by observation and that such creations of the imagination as are from time to time offered as "geotectonic hypotheses" have nothing in common with science.

To show that geotectonics has achieved considerable success in discovering the laws of the structural development of the earth's crust, so that knowledge of the principles of geotectonics has become a powerful tool in solving extremely diverse geologic problems, both theoretical and practical.

To show that geotectonics has its own program of study and employs its own particular methods of investigation, which give it the status of a fully independent scientific discipline.

To show the unity in diverse and contradictory tectonic processes, to find their interrelationships, and to discover the laws of their history and their development in space.

This book has a history of its own. The author's first theoretical book was published in 1948, with the title, *General Geotectonics*. The rapid development of this science required the publication, in 1954, of a considerably revised book on this subject, under the new title *Basic Problems in Geotectonics*. The latter book has been translated into Chinese and Japanese.

The book now before the reader is not merely an English translation of *Basic Problems in Geotectonics*. For the English edition, the author has greatly revised and brought up to date the whole text of his book. The part on tectonic movements that produce folds has undergone

particularly great changes. Important changes and additions have also been introduced into the parts devoted to the internal structure of the earth, to tectonic movements producing ruptures, and to the tectonics of the oceans. The entire text has been reedited. All these changes have been made with the aim of bringing the book as much as possible up to date by drawing on present-day concepts and data in the various fields of geotectonics.

The author has formed his opinions mainly from the enormous body of investigations carried out over the past ten years by the large company of Soviet geologists. He has nevertheless tried to make the greatest possible use of material from other countries. In his approach to the facts under consideration, in his attempts to see a unity in the processes that have produced tectonic structures and in his determination of the laws that relate these processes to each other, the author has undoubtedly been influenced by his teacher, M. M. Tetyayev, who must, in all justice, be ranked as one of the greatest modern thinkers in geology.

This book makes no compromises. The author analyzes the facts by the methods that he believes to be best founded. He draws the conclusions he considers to be the most justified, nowhere attempting to force them into agreement with the conclusions of other writers. In some cases his conclusions will be the same as those of others; at other times they will be sharply opposed. The author would like to see proof of the rightness of his views in the objective quality of the materials he has drawn upon and the methods he has used and in the fact that, in the light of the views he has presented, a unity actually does emerge from the diversity of tectonic phenomena.

This book is concerned primarily with problems of general tectonics. It does, however, briefly consider some of the data of structural geology and present a very rapid survey of regional tectonics.

The author considers himself greatly honored in appearing before the large and highly qualified audience of specialists and students who will read his work in the English language.

V. V. BELOUSSOV

PREFACE TO ENGLISH EDITION

Once every decade or so the geologic fraternity is treated to a monumental synthesis of earth structure. This is such a book, in the grand manner of Heim, Suess, Bucher, van Bemmelen, and other giants of the field. Like the books of many of his peers, Beloussov's text is dogmatic, overpowering in scope, variable in quality, and sparkling with gems of geologic insight. More importantly, it is the culmination of a period of vigorous scientific work in a society essentially closed to the West. As a consequence, the very lines of approach to geotectonic problems are unfamiliar, or at least rather neglected here.

Although for the most part a direct translation of *Basic Problems in Geotectonics*, published in 1954, parts of this book, especially the last third, have been extensively revised and amplified and new chapters added. The result, though very worthwhile, makes for unevenness of presentation. Accordingly, a brief statement of Beloussov's principal theses will be helpful to the reader unfamiliar with the work on which this book is based.

The theory of geotectonics herein presented—and it is essentially one unified theory—springs from the common geologic observation that the earth's crust has undergone oscillatory vertical movements, in which subsidence and uplift follow each other in more or less continuous rhythmic succession. Most parts of the crust available to study indicate a several- to manyfold repetition of the cycle. Evidence is to be found in the thickness of sedimentary layers and in the unconformities that separate the various layers.

Professor Beloussov assumes a degree of correspondence between subsidence and sediment thickness which will seem unlikely to American geologists but which is nevertheless logical if his initial premise is accepted. This premise is that each increment of subsidence is represented by an equal increment of sediments and hence by an equal amount of uplift in adjacent areas supplying the sediments. There are, of course, complications, but this in essence is the basic element for geotectonic analysis.

Three tectonic environments are recognized: geosynclines, platforms, and oceanic areas. Oceanic areas are little understood at present. Oscillatory movements characterize both platforms and geosynclines, but they are stronger and more sharply defined in the geosynclines. On the platforms, areas of subsidence and areas of uplift are named, respec-

tively, synclises and anteklises, and the movements tend to be long-continued, though intermittent, and to cover broad areas. In the geosynclines, on the other hand, an early period of subsidence culminates in a period of inversion or uplift; a central uplift with subsidiary depressions develops and gradually expands to occupy most of the original geosyncline.

Several stages in the formation of a geosyncline and its ultimate conversion to a platform are reflected in the sequence of sedimentary rocks. The geosynclinal cycle starts with a lower terrigenous sequence, which is followed by a limestone sequence marking the end of the first half of the cycle; this is the period of maximum dominance of subsidence over uplift within the geosyncline. The second stage, linked to initiation of central uplifts within the geosyncline, consists of clastic rocks and is designated the upper terrigenous sequence. A lagoonal sequence may follow if the climatic conditions are favorable, or the upper terrigenous sequence may be overlaid directly by a molasse sequence, which is the final stage of geosynclinal development. At this time high mountain ranges are present within the geosyncline, and uplift greatly predominates over subsidence. The geosynclinal belt has now become a part of the platform, and platform sequences of sedimentary rocks may develop.

Intense folding, faulting and intrusion, and extensive sialic igneous activity characterize geosynclines. The igneous activity, also the most intense folding, are related to uplifted zones—geanticlines—within the geosyncline, and folds are said to be characteristically asymmetrical away from the adjoining depressions toward geanticlinal uplifts. Furthermore the author rules out crustal compression entirely and explains all folding and faulting as a secondary effect of the oscillatory vertical movements.

The earth is believed to be divided into segments by a number of great deep-seated primary faults that cut completely through the crust. Oscillatory vertical movements tend to be concentrated along the deep faults, and the linearity of mountain systems is attributed to the fact that they follow primary deep faults.

Nearly everywhere basement rocks beneath sediments are made up of sialic igneous rocks and deformed metamorphic rocks, which are characteristic of the interior zones of geosynclines. The author postulates that during Archean time geosynclinal conditions were essentially world-wide. In the course of time, larger and larger areas have become converted to platforms and the geosynclines have gradually been restricted to smaller intervening areas between platforms. According to Professor Belousov, platforms first became prominent during the Proterozoic era and expanded after each succeeding period of deformation. Thus, the Caledonian geosyncline in Europe occupied a wider zone than the Hercynian geosyncline, which was, in turn, further restricted in the

Alpine cycle. A part of the Caledonian geosyncline became Hercynian platform and a part of the Hercynian geosyncline became Alpine platform.

Although the ultimate cause of the more or less constant oscillatory movements of the earth's crust cannot at present be demonstrated, the author favors differentiation of the earth's interior as the probable driving mechanism. By a process of diffusion, sialic material segregates above more mafic material, presumably at several levels within the earth. Radioactive substances concentrated preferentially within the sialic material differentiate, causing local heating, expansion, and eventual mobilization and injection of sialic igneous rocks and extensive metamorphism and granitization of geosynclinal sediments. A final stage of extrusive activity dissipates heat; the uplift then contracts and subsides. During active expansion and uplift, sediments are extensively stretched above the uplift. On subsidence, the sediments presumably collapse into folds, thus producing the intense folding characteristic of the central uplifted zones of geosynclinal belts.

Broader oscillatory movements affecting major parts of continents and more restricted movements of parts of platforms are related to differentiation at depth, resulting in expansion and periodic upward movements of masses of sialic material as the buoyant forces reach critical values: expansion is followed by a period of cooling, contraction, and subsidence. The exact mechanics remain something of a mystery. Professor Belousov is well aware of the uncertainties involved; he advances this hypothesis primarily as a guide for future research.

The problem of the origin of the oceanic depressions is troublesome. The author notes that oceanic areas appear to be geologically young and suggests that they are formed by some process of basification of previously continental areas, subsiding as the underlying crust becomes more mafic and hence heavier. The conditions that would reverse the trend toward differentiation and cause "basification" cannot now be specified.

Difficulties of translation make it almost inevitable that the author's views have not always been accurately rendered and that mistakes have crept into the text. A further source of confusion undoubtedly results from the fact that much of the non-Soviet literature was examined in translation by the author. Few ideas can survive two translations unchanged and unmarred. We can only hope that no major misrepresentations have occurred.

The major part of this book, based directly on the work of the author and his colleagues, is vigorous, lucid, and stimulating. As examples, one might cite the discussion of crustal structure (Chap. 3), the morphology and genetic classification of folds and faults (Chaps. 6, 7, 19, 20, and 21), and particularly those chapters dealing with oscillatory movements and their reflection in the sedimentary record. Understandably, those parts of

the book depending heavily on Western European and American literature are the weakest. The literature available to the author was necessarily rather randomly selected and in large part outdated. In this category would be included much of the material concerning igneous phenomena, granitization and metamorphism, and thrust and strike-slip faulting. No criticism is intended by this comment, however. The author has struggled with an impossible mass of literature and has accomplished far more with "foreign" literature than any western geologist has attempted with the Soviet literature. The book is first and foremost a statement of the present status of geotectonic theory in the Soviet Union and a record of the path by which the theory evolved. As such it is an important contribution to geologic literature.

As an outgrowth of his hypothesis, Professor Belousov has developed a geotectonic classification and extended it world-wide, as shown on the colored map accompanying this book. Regardless of one's views as to the ultimate validity of the hypotheses on which this classification is based, the resulting map is a useful compilation, emphasizing as it does the major elements of the post-Precambrian mountain systems. The data for Eastern Europe and Asia are particularly important for western geologists, who have little chance to follow the literature on these areas.

Some parts of the book are somewhat repetitious and dogmatic. Quite obviously it has not been subjected to the criticism of peers, or if so, it has not been greatly modified by such criticism. This however, is a source of strength as well as weakness: synthesis and development of a theory of global dimensions require much generalization in areas peripheral to the main thesis, which can be accomplished only by a writer free to pick and choose the facts, interpretations, and suppositions that bolster his thesis.

JOHN C. MAXWELL

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This monograph was translated under the auspices of the American Geological Institute with financial assistance through a grant from the National Science Foundation. It was selected for translation in 1959 by the institute's Advisory Committee on Translations, Rhodes W. Fairbridge, chairman.

The American Geological Institute is indebted to the author, Professor V. V. Belousov, for his assistance in the review of the translated manuscript. Numerous corrections and additions to the original book were made by the author, and he rewrote several chapters, incorporating extensive revisions.

Professor John C. Maxwell, of Princeton University, served as the scientific editor of the English translation. Dr. Maxwell has written the

foreword to the translation. Staff supervision was provided by Martin Russell and Thomas F. Rafter, Jr., of the American Geological Institute.

The American Geological Institute is a federation of scientific and technical societies in the area of the geologic sciences and operates under the National Academy of Sciences—National Research Council.

BIOGRAPHICAL SKETCH OF THE AUTHOR

Vladimir Vladimirovich Belousov was born in Moscow in 1907. In 1930 he finished his studies in the department of geology of the University of Moscow. From 1930 to 1941 he lived and worked in Leningrad. V. V. Belousov began his scientific work with a study of the geology and geochemistry of natural gases; however, he soon turned to the problem of geotectonics, which has since become his specialty.

He has been living in Moscow since 1942. He is director of the department of geodynamics in the Institute of Earth Physics, Academy of Science of the U.S.S.R., as well as being a professor at the State University of Moscow. His scientific interests include the mechanism and laws of tectonic deformation, tectonophysics, experimental tectonics, and general problems of geotectonics and geophysics. He is presenting an advanced course in geotectonics at the University of Moscow.

V. V. Belousov has done his field work in tectonics, both currently and in the past, in the Caucasus; he has also worked in the Crimea, the Urals, the Transbaikal, in Tien Shan, and in the Kara-Tau. In 1947 he worked in the Eastern Alps. In 1955 he led geologic expeditions in China and in the French Alps.

In 1953 he was elected corresponding member of the Academy of Science of the U.S.S.R. He is also a member of the Geological Society of France and a corresponding member of the Geological Society of America.

In 1957 he was elected vice-president of the International Union of Geodesy and Geophysics. As vice-president of the Soviet Committee for the I. G. Y., he participated actively in organizing the work of the International Geophysical Year. He is also a member of the governing board of the Special Committee (International) for the I. G. Y.

At the Twelfth General Assembly of the International Union of Geodesy and Geophysics held at Helsinki, Finland, in 1960, V. V. Belousov was elected its president.

V. V. Belousov's main scientific publications are as follows:

1. *Problems of the Geology of Helium*, 1933.
2. *Mud Volcanoes of the Kerch-Taman Region: The Conditions of Their Formation and Their Activity*, 1936.

3. *Outline of the Geochemistry of Natural Gases*, 1937.
4. *The Greater Caucasus: Results of Investigations in Geotectonics*, 1938–1940.
5. *Facies and Thicknesses of the Sedimentary Series in the European Part of the U.S.S.R.*, 1944.
6. *General Geotectonics*, 1948.
7. *Tectonic Ruptures, Their Types, and the Mechanism of Their Formation*, 1952.
8. *Basic Problems in Geotectonics*, 1954.
9. *The Geologic Structure and Development of the Ocean Basins*, 1955.
10. *The Types and Origins of Folding*, 1958.

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PART ONE

GEOTECTONICS AS A SCIENCE:
THE METHODS AND HISTORICAL DEVELOPMENT
OF GEOTECTONICS

CHAPTER 1

GEOTECTONICS AS A SCIENCE

THE SUBJECT AND PROBLEMS OF GEOTECTONICS

The word "geotectonics" is derived from two ancient Greek words: *ge*, "the earth," and *tektonike*, "building." This branch of science is sometimes called simply "tectonics," but some writers tend to use "tectonics" to refer only to the morphologic aspects of geotectonics—the description of rock structure—and to use "geotectonics" for more general problems.

Geotectonics is a branch of geology; along with the other branches of this science, it is concerned with the structure and development of the earth's crust. The subjects of geotectonic research are those features of the structure and development of the earth's crust which are produced by mechanical processes arising from forces deep within the earth. Geotectonics determines the laws that govern these processes and the causes that produce them. This definition, however, must be clarified. We must explain what is meant by "the structure of the earth's crust" and by "mechanical processes arising from forces deep within the earth."

We find that the distribution of materials in the earth's crust is not uniform. The rocks composing the earth's crust are different in different places. Each rock forms a geometric body within the earth. In some cases rocks occur in strata and change vertically from bed to bed. In other cases rocks form massive bodies of various shapes; these may be cylindrical, ellipsoid, mushroom-shaped, etc. Rock strata may occur horizontally, but they may also be inclined, bent into folds, or crossed by fractures. This lack of uniformity in the distribution of the rocks within the crust will here be called the *structure of the earth's crust*.

By *mechanical processes* are meant the movements of the earth's crust and its deformations. It is known that the crust of the earth is in motion, rising up in some places and subsiding in others. It is tilted and bent, and various parts of it are crumpled into folds or broken by fractures.

Thus geotectonics includes the features of the earth's crust that are produced by its movements and deformations but not the chemical or

physicochemical processes within the crust, even though these lead to the formation of different types of rock and thus affect the structure of the earth's crust. Geotectonics thus may be called the *mechanics of the earth's crust*. The problems of geotectonics are somewhat broader, however, in that this branch of geology, although it does not concern itself with the essence of the physicochemical processes in the earth's crust, does establish certain laws that govern the manner in which these processes develop, inasmuch as they are associated with movements and deformations of the crust. Igneous phenomena as such, for example, since they are fundamentally physicochemical, are not specifically considered in geotectonics. But igneous activity is studied to the extent that it suggests mechanisms that associate the formation of igneous rocks with mechanical movements of the earth's crust. Geotectonics examines the formation of igneous bodies within the earth's crust, since the shape and distribution of these bodies are in some measure determined by displacements of magmatic material and the need to make room for it by fractures and other means.

The movements and deformations of the earth's crust primarily affect the mode of occurrence of rocks. The location of minerals in rocks is determined initially by other processes, which fall within the scope of petrology or lithology. Thus geotectonics studies mainly the structure and occurrence of the rocks themselves within the crust of the earth. But in a number of cases mechanical deformations of the crust also affect the orientation of individual minerals in the rock. For example, minerals that were earlier distributed at random may, under the influence of deformations, acquire a regular orientation. Such cases, naturally, lie within the scope of geotectonics.

Mechanical processes in the earth's crust may have either internal or external causes; geotectonics studies only the first. It is not concerned with exogenic processes of weathering and rock formation, avalanches, landslides, and karst phenomena, although all these to some extent affect the structure of the outer layers of the earth's crust.

It is not always easy to draw the line between internal and external (endogenic and exogenic) mechanical processes within the crust. In some cases, moreover, it is hard to distinguish tectonic from nontectonic processes. For example, there is no doubt that the force of gravity plays a large role in certain processes that are known to be tectonic; on the other hand, gravity is also the cause of many nontectonic processes, such as landslides and avalanches. An exact boundary between tectonic and nontectonic processes that change the earth's structure is also difficult to draw because of the scale of the phenomena: a fault that is definitely tectonic may in a given case be much smaller than a large landslide. At the present level of knowledge it is not possible to find a single, precise criterion by which one may in all cases distinguish tec-

tonic processes, as well as tectonic structures of rocks, from nontectonic. Therefore, to avoid artificial formulations that will later prove to be sterile, tectonic movements and tectonic structures will be limited to those associated with displacements of the material in the earth caused by internal forces. These forces include the force of gravity and specifically exclude such processes as, and such occurrences as are known to be caused by, avalanches, landslides, glaciers, weathering, the action of surface and ground waters and the wind, and also the activities of man.

Is geotectonics concerned only with the earth's crust, or with the earth as a whole? It is devoted primarily to problems connected with the structure and history of the earth's crust. Since it is a geologic discipline, and thus based chiefly on direct study of rocks and the conditions under which they occur, geotectonics as such offers very limited possibilities of penetrating into the depths of the earth below the crust. A study of the structure and development of the earth's crust provides a sufficient basis for solving most practical problems in present-day geotectonics. In dealing with general problems involving the causes of movements and deformations of the earth's crust and the laws governing the development of large parts of it, however, one cannot consider the crust alone, apart from the deeper regions of the earth. The formation of crustal structure is very closely linked to processes that arise in the depths, and the history of the earth's crust is only one part of the history of the planet as a whole.

In dealing with these general problems, geotectonics merges with geophysics and geochemistry, using the data and methods of these sciences for its own ends. From the combination of geotectonics, geophysics, and geochemistry there should ultimately arise a general science of the earth.

To achieve its ultimate goal of understanding the structure of the earth's crust and the mechanical processes that take place within it, geotectonics must successively carry out a number of separate tasks that are closely linked to each other.

The first task is to study the distribution of the rocks in the earth's crust. The geometric bodies formed by the rocks in the earth are the modes of occurrence, or structures, of the rocks. These include beds, folds, batholiths, laccoliths, and veins. Thus the first task includes studying the structural forms in which the rocks occur; in carrying out this task the geologist must obviously define and classify the various structures he observes and describe their characteristics. The division of dislocations into folds and fractures and the establishment of various types of folds (linear, domical, symmetrical, asymmetrical, recumbent, box folds, etc.) are examples of such classifications.

The second task is to study and describe the present distribution of

the various structures in the earth's crust. Here the structural features peculiar to each individual region may be listed. Some, for example, are characterized by the occurrence of horizontal beds of a particular age and others by typically elongated linear folds; in still others the beds of the same age are crumpled into box-shaped folds.

The third task is to determine the regularities in the horizontal and vertical distribution of the structural forms. Repeated observations, for example, have established that regions of intensive linear folding grade into regions of horizontally lying beds (platforms) through intermediate zones characterized by the development of domical dislocations. This general rule has very few exceptions and, consequently, may be called a structural law. Similarly, other laws of structure include the occurrence of overturned folds inclined toward the platform at the margins of folded zones, the occurrence of batholiths in the interior of folded regions, and the close relationship between the shapes of folds and the distribution of fractures in them. At the present time geotectonics has established a number of such regularities. Knowledge of these regularities and skill in using them are of great importance in practical work.

The fourth task of geotectonics is to study the history of the observed structures and thus explain the nature and the direction of the tectonic movements that have led to the occurrence of one structural form or another. Any object of interest to geology is studied in such a way as to discover its history and the conditions under which it was formed. Since one can rarely observe geologic changes directly and can see only the results of these changes, one must try to detect traces of earlier changes and movements of the materials in the earth. Similarly, geotectonics attempts to find traces of previous changes by observing structural forms and, by this means, to explain their history. For example, in observing the folded occurrence of a bed, we know that the entire bed was originally laid down horizontally. It follows that after the bed was deposited, there was a dislocation in which the material in this region was displaced from its original position and the strata were bent. By comparing the observed occurrence of the beds with their inferred original horizontal position, one can determine the nature and the direction of the displacement.

By such observations and inferences one may establish definite criteria for recognizing former changes in structural forms, along with displacements of the material in the earth. Such displacements are called *tectonic movements*; these are understood as mechanical displacements of the material of the earth's crust producing changes in the structural forms, or, in other words, changes in the mode of occurrence of the rocks.

As a result of the solution of the above-mentioned problems, the earth's crust may be divided into various structural regions, or structural

complexes, as they are often called, and into regions with different structural histories. A region that has a particular structural environment, certain predominant structural forms, and a history different from that of other regions is called a *structural*, or *geotectonic*, *zone*. The establishment of such zones is the task of geotectonic regionalization. A very simple example is the division of the earth's crust into folded zones and platforms.

The fifth task is to discover the laws governing the succession and interrelationships of tectonic movements. For example, it has been established that folding is generally preceded by an intense subsidence of the earth's crust and the formation of a geosyncline. It has also been established that granite intrusions are formed, for the most part, at the same time as the main phase of folding, whereas conformable lava flows and sills precede it. It is known that the history of the earth's crust is characterized by structural changes in definite directions and by periodic recurrences of similar tectonic phenomena. By solving this fifth problem, the structural geologist may establish the general sequence of tectonic movements and explain the relationships between them.

The sixth task is to generalize. It consists in elaborating a geotectonic theory, that is, explaining geotectonic phenomena in all their complexity and the causes and forces that produce them. As we shall see, such a general theory has yet to be worked out; meanwhile, we are forced to use a number of partial theories to throw light on the relationships between various types of tectonic phenomena. To explain tectonic processes as a whole, there are only a number of hypotheses.

THE BRANCHES OF GEOTECTONICS

Geotectonics may be divided into a number of branches corresponding to the tasks enumerated above. The study of the structural forms of rocks and the classification of these structures are the subject of morphologic geotectonics or structural geology. The description of the regional distribution of structural forms and the study of the structural history of particular regions or of the earth's crust as a whole are included in regional geotectonics. The remaining problems—the regularities in the horizontal and vertical distribution of structural forms, their history, and the laws governing their development, as well as the elaboration of geotectonic hypotheses and theories—lie within the domain of general geotectonics.

All these branches of geotectonics are closely interwoven, so that it is often difficult to determine the boundaries between them. The most clearly circumscribed task is that of structural geology. The establishment and classification of the types of structural forms are both the beginning and the basis of every geotectonic investigation. Regional geo-

tectonics integrates the data, establishes the horizontal and vertical distribution of structures, and studies the history of their development. General geotectonics uses the results of these branches of geotectonics, analyzes them, generalizes, and draws conclusions regarding the laws and mechanisms in the development of structural forms. General geotectonics, in the final analysis, produces universal geotectonic concepts by generalizing from the data of regional geotectonics for the entire surface of the earth, supported by data from geophysics, geochemistry, and other related sciences. General geotectonics, in turn, may be divided into (1) historical geotectonics, which studies the history of tectonic movements and transformations; (2) the dynamics of the earth's crust or of the earth (geodynamics), which investigates the physical conditions of tectonic movements and transformations; and (3) theoretical geotectonics, which is concerned with the creation of geotectonic theory.

This book will deal with the fundamental problems of general geotectonics. It is assumed that the reader has become familiar with the fundamentals of structural geology from other sources, although some of these are briefly presented here. The author has introduced a fair amount of material from the field of regional geotectonics for the purpose of illustration, but the regional geotectonic survey of the earth given in this book is of necessity very short.

THE POSITION OF GEOTECTONICS AMONG OTHER BRANCHES OF GEOLOGY

Geotectonics occupies a special and very important place among the various branches of geology. In pursuing its goal of discovering the structural history of the earth's crust and establishing the laws and the forces that direct its development, geotectonics generalizes from a number of branches of geology. This generalizing importance of geotectonics will become evident later; here we may illustrate it with a few examples.

The structural geologist studies facies, along with the history of the oscillatory (epeirogenic) movements of the earth's crust, to draw conclusions regarding the influence of tectonic movements on the formation of sedimentary rocks. He also uses geomorphology to relate the development of land forms to the same oscillations. On the basis of data from petrology, he explains the relationship of various manifestations of magmatic activity to specific structures as well as to tectonic movements and the stages in the development of the structure of the earth's crust. By comparing metamorphism with tectonic movements, he determines the laws governing the manifestation of various kinds of dynamic metamorphism; that is, metamorphism caused primarily by mechanical processes.

As will become clear later, the facies and thicknesses of sedimentary rock suites, their changes in time and space, and the land forms on the earth's surface are, along with the development of structural forms, determined fundamentally by tectonic movements, whose cumulative effect may be called the *tectonic character* of a region. Discovery of the laws of the development of tectonic movements and of their causes in time and space leads to the establishment of a number of laws of the formation of sedimentary rocks and the relief of the earth's surface. Thus one can clearly see the interrelationship between geotectonics and other branches of geology. By using the data of the other branches of geology to derive its own generalizations and unifying these results to some extent, geotectonics provides a general theoretical framework for these branches and also helps interpret their data and suggest directions for further research.

The study of actual structures is the basis of every geotectonic investigation; this is done by drawing geologic maps and sections. Thus geologic mapping is widely used in geotectonics as the means of obtaining basic factual material. In order to construct a good geologic map, on the other hand, one must know the geometry of structural forms and the laws governing their distribution as established by geotectonics. Stratigraphic data are indispensable in constructing geologic maps and sections and in establishing the geometry of structural forms. In finding the relative age of rocks, one must determine their original distribution in space, such as, for example, the sequence of their stratification. This must be known, in turn, in order to understand the present disposition of the rocks, which has been altered by tectonic movements. Furthermore, stratigraphic data are necessary for reconstructing the history of tectonic processes.

Geotectonics, in its turn, helps solve many problems in geochronology, since tectonic movements determine the rhythm of the earth's history and the boundaries between stratigraphic subdivisions generally correspond to phases of these movements.

The structural geologist must use the results of historical geology, since by analyzing the changes in the sea and the contours of the dry land, and in the facies of sediments, he can reconstruct the history of oscillatory and other movements of the earth's crust. On the other hand, the history of past geologic epochs and their changes in physical geography can be properly understood only through a knowledge of the laws governing the development of tectonic movements. In addition, in using the methods of paleogeography, geotectonics perfects these methods and establishes the sequence and interrelationships of the phenomena studied by historical geology.

In studying the facies and the petrography of sedimentary rocks, ancient sediments are compared with those of today, and the physical

and geographic conditions under which they were formed are determined from their lithologic composition and structure. Geotectonics uses such reconstructions of the manner in which physical geography has changed in time and space to establish the connection between these changes and tectonic movements. Since these movements follow certain laws, these various branches of geology are given a theoretical foundation that facilitates investigations in their own proper fields.

Through a knowledge of data supplied by petrology on the physico-chemical conditions under which igneous bodies were formed, one can understand the mechanical conditions -pressure, temperature, jointing, etc. in the earth's crust in a given geologic epoch. By establishing the connection between various types of magmatic activity and the movements of the earth's crust, geotectonics helps the petrologist determine the mechanical conditions under which intrusive and extrusive igneous rocks have been formed.

The connection between geotectonics and geomorphology has already been indicated. Geomorphologic analysis is a very important tool in studying recent tectonic movements. But a proper understanding of land forms must be based on a knowledge of the laws of vertical movements of the earth's crust.

The study of mineral deposits, both ore minerals and others, is the source of valuable data for geotectonics, which helps determine the association of various mineral deposits with rock formations and structures in the earth's crust. By discovering the fundamental laws governing the creation of these rock formations and structural forms, geotectonics, in turn, aids in the solution of problems in applied geology. Metallogenetic provinces and belts and coal- and oil-bearing zones are closely linked with regions of a definite tectonic character. Although such connections are still not completely understood, they are nevertheless of practical value as a tool in predicting the distribution of mineral deposits.

Engineering geology provides interesting material for geotectonics in its experimental studies of the mechanics of the earth's crust. Data on, for example, the strength and rigidity of rocks, the distribution of stresses in rocks, their stability in the roof of a subterranean gallery, or their plasticity, obtained from engineering geology, are a great help in solving mechanical problems in geotectonics. Geotectonics, in turn, by establishing the structural forms of rocks, the development of fractures, etc., facilitates the solution of problems in engineering geology.

Geotectonics makes extensive use of geophysics in discovering the structure of the earth's depths; geophysics, on the other hand, must use geotectonic data, since geophysical concepts might otherwise lose touch with geologic facts.

Recent times have seen manifold connections between geotectonics

and geochemistry. It is very likely that the mechanical processes in geotectonics are based on more general processes involved in the geochemical transformations of materials in the earth. This set of problems is just beginning to be touched upon.

Geotectonics also comes into contact with other sciences outside the discipline of geology. A closer connection must be established between geotectonics and cosmogony, since the origin of the earth, as revealed by cosmogony, must in great measure have determined the course of its structural history. Comparative study of the earth, the moon, and the other planets may shed a great deal of light on this question. There are, naturally, great differences in the external appearances of the planets, although all planets evidently have a similar origin; such differences can be clearly seen by comparing, for example, the earth and the moon. The absence on the moon of elongated linear mountain ranges and the presence of huge volcanic craters and vast circular depressions ("seas"), which appear to be filled with volcanic rocks, suggest numerous problems to be investigated. A comparison of the surfaces of the earth and the moon may prove invaluable for understanding tectonic processes that take place on our planet. Such a study might bring in Mars and its "canals," which may be tectonic cracks on the surface of that planet. On the other hand, geotectonics may be of help to astronomers in, among other ways, solving problems of the evolution of the planets and in explaining what has been observed of the relief of their surfaces.

A connection exists, or could be established, between geotectonics and certain branches of physics, such as the study of plastic and elastic deformations. Data from these fields of physics will help the structural geologist to understand the mechanisms in the deformations of the earth's crust. Geotectonics, in turn, will provide physics with examples of deformations in which time has played an exceptionally important part—a factor that cannot be reproduced under experimental conditions. Geotectonics also provides examples of the deformation of bodies with enormous volumes. Very recently a separate branch of science known as *tectonophysics* has emerged from the union of geotectonics and physics.

GEOTECTONICS AS AN INDEPENDENT BRANCH OF GEOLOGY

Geotectonics was developed relatively late as compared with many other branches of geology. The word "geotectonics" itself first appeared in 1850, but this branch of knowledge has developed into an independent scientific discipline on a level with the others in the family of geologic sciences only since 1930.

Geotectonics has a proper place as an independent branch of geology, first of all, because of the specific nature of its problems, which have been enumerated above. These problems as a whole do not lie within

the scope of any other branch of geology, and their extent is very great. In the second place, the independent status of geotectonics is justified by its peculiar approach to these problems: structural geologists consider all geologic phenomena as the direct or indirect result of movements of the earth's crust. The independence of geotectonics, finally, is based on its own special methods of investigation. Although the basis of these methods is derived from other branches of geology, geotectonics modifies and develops them in accordance with its specific approach to the phenomena it studies.

THE CHIEF METHODS OF GEOTECTONICS

1. *The structural method.* This consists of a study of the modes of occurrence of rocks and their internal texture and structures. The geologist is concerned with structural forms of various scales and orders; for convenience, these may be divided into large, medium, and small structures.

Large structures are such structural complexes as will be revealed by regional geologic mapping: large folds, faults, and large igneous bodies such as batholiths, laccoliths, and stocks. By small structures are meant rock textures of tectonic origin. Such textures are observed in so-called tectonites—rocks that have been modified internally as a result of tectonic changes. The most important feature in the texture of tectonites, which will be discussed later, is the regular orientation of minerals, which may be manifested, for example, as schistosity. The various types of mineral orientation are small structures. Between these two extreme groups is a group of intermediate structural forms. This group includes all structures that can be observed macroscopically but are of such small dimensions that they are not, as a rule, recorded in regional geologic surveys. Small folds and crenulations, large complex folds, small ruptures with displacements of several meters' amplitude, joints, cleavage and schistosity, and small igneous bodies and hydrothermal veins, etc., all belong to this group.

It is quite obvious that structural forms of every order must be studied to obtain a complete and correct idea of the structure and development of the earth's crust. The tectonist uses data on the morphology of large structures, which he obtains by geologic mapping, and analyzes these with a view to explaining the history of such structures. From the morphology of the structures, the geologist infers the movement of the material in the earth's crust and its physical state. By the shape of folds, for example, it can be determined whether they were formed by plastic horizontal or inclined flow and whether the rock was in a state of high or low plasticity. The direction of flow may be obtained from the inclination of the axial plane. If the degree of plasticity

of the material at the time of deformation is established at the same time as the direction of flow, many additional problems may be solved. It is possible, for example, to interpret the character of visible but poorly exposed faults that accompany folding: faults of which the strike and distribution are closely connected with inclined isoclinal folds most probably represent imbricated overthrusts with displacement toward the slope of the fold; faults that cut domical uplifts, the shape of which indicates they were created by vertical forces, are probably normal or reverse faults.

In studying large structures, geotectonics uses, for the most part, the methods of ordinary geologic mapping. Medium and small structures, however, can be fully investigated only by using specifically geotectonic methods. The study of medium structures involves detailed mapping of small structural forms on a large scale and statistical measurement of the positions of joints, cleavage, and schistosity. From these statistical methods special diagrams can be drawn to visualize the deformation: the positions of the stress axes, the direction of movement of the material, etc. From the plan of the deformation it is possible to infer the mechanism of formation and the history of the given medium structure.

The essence of studying small structures is to determine the positions of the minerals in the rocks in relation to the elements of the medium and large structure. This is done by finding the orientation of minerals relative to the position of the rock (for example, the arrangement of elongated minerals with their long axes in one direction) and the optical orientation of the minerals (the position of the optical axes). Statistical methods of investigation are of great importance here. Study of the small structures also reveals the plan and the nature of the deformation, its separate stages, and the physical state of the material at the time of the deformation.

A study of medium and small structures produces a much more complete understanding of the mechanism and history of a structural form, since by this means it is possible to see the details of the tectonic movement that has taken place in the rock.

2. *The investigation of oscillatory movements of the earth's crust by studying the thicknesses and facies of sedimentary rocks.* This method has to a large degree been worked out by Soviet geologists: it is based on the distribution of the thicknesses and facies of sediments in space and time. The thickness of the sediments indicates the amplitude of the oscillatory movements of the earth's crust. By comparing the distribution of large and small thicknesses of sediments within various stratigraphic subdivisions, a fairly complete history of the slow vertical movements of the earth's crust may be reconstructed. The facies of sedimentary rocks are studied to determine the location during various epochs of areas of erosion, of dry land and water, and of deep and shallow seas.

Comparison of the results of these studies, like the analysis of thicknesses, makes it possible to reconstruct the history of the oscillatory movements of the earth's crust.

It should be noted that the distribution of sedimentary facies is also of concern to other branches of geology, such as paleogeography or the special study of facies as such. In these branches, however, facies are studied primarily to explain the physical geography of previous epochs; tectonists, on the other hand, study the distribution of facies and thicknesses in order to reproduce the history of the oscillatory movements of the earth's crust.

3. *The volumetric method of studying the oscillatory movements of the earth's crust.* This method, which was developed by A. B. Ronov, borders on the previous one, since it is based on calculation of the volumes of sedimentary deposits. The total volume of sediments accumulated during any given length of time is approximately equal to the total subsidence of the earth's crust during that time. The volume of terrigenous deposits corresponds to the "volume" of the uplifts that were the source of clastic material for these deposits. Hence, by comparing the total volume of sediments accumulated in a given territory over a given stretch of time with the volume of terrigenous material in this territory, the relationship between the uplift and subsidence of the earth's crust can be obtained. A change in this relationship with time, furthermore, will explain various aspects of the development of oscillatory movements of the earth's crust.

4. *The geodesic method.* This method of studying present-day movements of the earth's crust is based on a comparison of geodesic measurements made (usually by leveling) at intervals of several years. From such comparisons it is possible to detect changes in the relief of the earth's surface, that is, movements of the earth's crust. Here the specific nature of the structural geologist's approach to his subject is especially clear: the techniques of geodesic measurement have no relation to geotectonics, which uses these techniques to solve its own specific problems.

5. *The geomorphologic method.* This method is used in studying present-day and relatively recent (Quaternary) movements of the crust of the earth. Its basis is the connection between the movements of the earth's crust and the relief of its surface, by which one may establish the character of these movements from the topography. A study of the shapes of valleys, terraces, ancient levels of erosion, and other elements of relief is part of this method.

6. *Geophysical methods.* Only the uppermost parts of the earth's crust can be observed directly; nevertheless, some idea of the structure of the deeper parts of the earth's crust is necessary for even a partial solution of geotectonic problems. Penetration into the depths of the earth is especially important in solving more general geotectonic prob-

lems involving movements of large portions of the earth's crust and the causes of such movements. The structure of the deep-seated parts of the earth, which are hidden from direct observation, is studied by various geophysical methods: seismic, gravimetric, magnetic, and electrical. The data obtained by these methods are indispensable in present-day geotectonics.

The usefulness of geophysical methods is not limited, however, to bringing out the structure of the deeper parts of the earth; some of these methods can also be used in studying present-day tectonic movements.

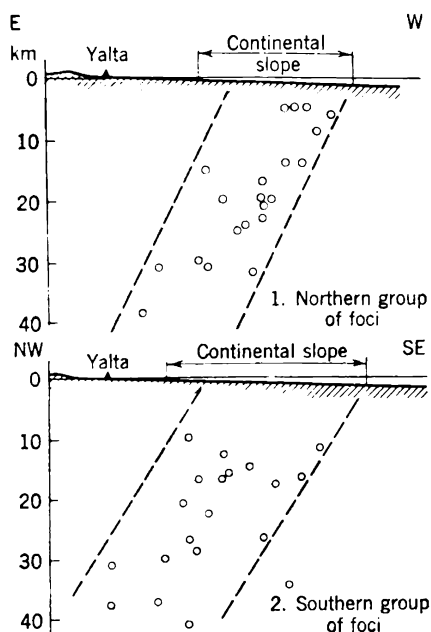


FIG. 1. Distribution of earthquake foci in the vicinity of the southern shore of the Crimea (section—after G. P. Gorshkov and A. Ya. Levitskaya). Horizontal and vertical scales are the same.

From seismic observations one may determine the location of earthquake foci and discover the mechanism of the movements they produce. Thus one can find the location of present-day “active” faults in the earth's crust, which determine the occurrence of displacements under the influence of tectonic forces in modern times.

For example, a study of the distribution of the foci of earthquakes occurring in the Crimea has shown that these are concentrated in a zone dipping steeply under the southern shore of the Crimea (Fig. 1). Thus one can mark the location of a fault zone separating the Crimea from the Black Sea basin; these faults have the character of steep reverse faults, which are still active (Gorshkov and Levitskaya, 1947).

Recently a number of attempts have been made to include observations of changes in the force of gravity with time at particular points among the methods of studying present-day movements of the earth's crust. Geophysical and, in part, astronomical methods are also beginning to be used to determine the distribution of various densities within the earth and the changes in this distribution over the course of time. New observations of the nonuniform rotation of the earth, for example, have shown that there are abrupt changes from time to time in the density of the material within the earth (Pariyskiy, 1948).

7. *Laboratory models.* At times various investigators have resorted to experimental reproduction of particular tectonic processes in order to obtain a clearer picture of them and to study the conditions under which they develop. For example, in boxes with one movable wall layers of various plastic materials (wax, paraffin, clay) have been subjected to horizontal compression, in imitation of the strata in the earth's crust. Folds of various shapes have been obtained, depending on the composition of these synthetic strata, the rate of deformation, and so forth. From the results of these experiments conclusions have been drawn regarding the conditions under which folds are formed in nature.

The method of using scale models is of great value in geotectonics, and the prospects for its future development are favorable. A great amount of preliminary work must be done, however, if this method is to be applied correctly. Many tectonic experiments have been carried out without meeting the requirements of the theory of physical similarity. According to this theory, a decrease in the dimensions of the bodies to be deformed, the effective forces, and the duration of the process must, if the similarity to natural conditions is to be preserved, be accompanied by a proportional decrease in the viscosity of the materials. Rough calculations show that, under laboratory conditions, heavy petroleum asphalt must be used instead of the rock salt that forms the nucleus of diapir domes. Many rocks can be replaced by petrolatum.

It has been shown that in the scale-model reproduction of slow plastic deformations, in which neither inertia nor elasticity is considered, the condition for the similarity takes the form

$$C_h = C_p C_e C_t$$

where C_h = the ratio of the viscosity of the model to that of the object being modeled

C_p = the ratio of the model's density to that of the natural object

C_e = the ratio of the model's geometric dimensions to those of the natural object

C_t = the ratio between the durations of the process in the model and in nature

(V. V. Belousov, 1958; M. V. Gzovskiy, 1954, 1958; Hubbert, 1937; Ye. N. Lyustikh, 1949; Shneyerson, 1947).

Below, in considering the mechanism of folding and fracturing, extensive use will be made of data obtained from such models, the great value of which is that one can follow the steps in the development of processes that appear in nature only in the form of final results.

To use models properly, one must know the properties of rocks under the temperatures and pressures within the earth's crust. For this purpose one must make laboratory experiments under high pressures; such investigations, however, have not yet been made on a sufficient scale. To understand correctly the results of experiments with models, one must also make an appropriate physical analysis of the deformation being reproduced. A theory of these deformations must also be developed. These problems together make up the subject of a new branch of science on the border between geotectonics and physics, which has been named *tectonophysics*. Tectonophysics studies the mechanics of tectonic processes, using detailed investigations in the field in combination with laboratory models and theoretical analysis. In the U.S.S.R. in recent years the systematic development of tectonophysics has begun.

THE PRACTICAL AND THEORETICAL IMPORTANCE OF GEOTECTONICS

The basic practical purpose of geotectonics is to explain and predict the laws governing the location of rocks and the structures formed by them in the earth's crust. This is necessary in drawing geologic maps and sections and in searching and prospecting for mineral deposits.

As a rule, the geologist cannot see directly the complete sequence of rocks in the upper layers of the earth's crust. Thus he can draw accurate conclusions only if he knows how structures are formed and how they are interrelated and distributed. This knowledge is provided by geotectonics. Geotectonics also makes it possible to predict the location of rocks and their depths below the earth's surface; this is extremely important in predicting the distribution of mineral deposits.

Observations have shown that there is a connection between geochemical and geotectonic processes. It is known that deposits of various metals are found in folded zones of different ages and structures. Although this association at the present time is empirical and has still not been fully explained, it is extensively used in predicting the location of economic minerals. An accurate estimate of the extent of a deposit can be made only with a knowledge of its structure, and this is obtained only with the help of geotectonics.

Without a proper understanding of the structure and tectonic history

of an area, it is impossible to divide it into seismic regions; that is, to predict the possible location of earthquakes of various degrees of violence. Such predictions are of great importance, since they are the basis of ensuring beforehand that buildings will have sufficient strength.

Geotectonics is extremely important in geologic theory. By explaining the laws and forces that determine tectonic movements of the earth's crust, geotectonics helps to illuminate the fundamental principles of the earth's history; in doing so, it touches on cosmogony. In addition, geotectonics provides the theoretical basis for solving problems of specific and general geologic significance. Even for specialists in the various branches of geology, the correctness or incorrectness of Wegener's hypothesis of continental drift, for example, is directly relevant. It is also not without significance whether vertical or horizontal forces predominate within the earth's crust. Various explanations might be suggested for the phenomena studied by these disciplines, depending on which point of view is the correct one. All concepts of the migration of flora and fauna and of the changes in the earth's climate, for example, will be affected by the correctness or incorrectness of Wegener's ideas.

CHAPTER 2

PRINCIPAL STAGES IN THE HISTORY OF GEOTECTONICS AND ITS PRESENT STATE

MAIN STAGES IN THE DEVELOPMENT OF GEOTECTONICS TO THE EIGHTEENTH CENTURY

Since there is no space here for a complete history of geotectonics, we shall discuss only the most important stages in its development. The development of geotectonics, like that of all geology, has always been determined by the use of mineral raw materials and by various technical problems whose solution involves penetrating the depths of the earth (the construction of canals and tunnels and the erection of large buildings). Practical requirements have always stimulated the growth of the geologic sciences. It may thus be said that scientific ideas in geotectonics began to appear mainly at a time when there was a significant development of the mining industry and an urgent need for understanding the structure of the earth's crust in the interest of exploration and prospecting. This did not happen before the eighteenth century. The earlier period must be considered the prescientific stage in the history of geotectonics.

Some of the first observations that are now part of geotectonics were made as long ago as the period of ancient Greek culture. Ancient thinkers, for example, were able to observe very early that the earth's surface does not remain immobile but undergoes certain movements and changes. In some places far from the shore, fishermen looking through the transparent sea water saw ruins of submerged buildings and whole cities on the bottom. In other places villages that were known to have stood on the seashore earlier were found to be inland at distances of several score kilometers from the shore.

There were also more penetrating observations. As early as the beginning of the fifth century B.C. the Greek philosopher Xenophanes, having discovered sea shells and vestiges of other marine organisms in rocks, drew the conclusion that at some time there had been sea where there was now dry land (Makovel'skiy, 1914).

The famous historian of antiquity Herodotus described instances of shifting of seashores, drying up of the sea bottom, and other changes in the earth's surface [Herodotus (Herodotus), 1888].

The ancient Greek geographer Strabo also cited many instances of such changes in the earth's surface in his *Geography* (Strabo, 1831–1833). He wrote: “. . . The sea rises and falls, it floods many places and retreats again, and the reason for this is that the bottom of the sea sometimes rises, and sometimes sinks, raising or lowering the water along with it” (Book I, page 78).

In his *Meteorology*, Aristotle said (Aristotle, 1863, page 86):

The same places on the earth are not always flooded or dry; their condition changes, depending on the appearance and disappearance of water. The relationship between the continent and the sea changes, therefore, and the same places are not always land or sea. The sea comes in where there was land before; land will return where today we see the sea.

As in other fields of science, these correct observations and intelligent arguments on the part of ancient authors are interwoven whimsically with fiction, fairy tales, and superstitions. Even in the most authoritative works of ancient scientists, one can easily find carefully recorded authentic facts side by side with legends. The fantasies become especially numerous when ancient authors speak of the causes that produce changes in the earth's surface. The majority of thinking persons of that time were apparently satisfied with myths as explanations for the phenomena of nature.

The Middle Ages, as we know, were a time of long stagnation in the sciences, including geology.

Geologic science began to be freed from the influence of church dogma in the sixteenth century, in the same period that witnessed the dissolution of feudal society and the rise of commercial capitalism. At this time there arose a demand for metal for coinage, for the manufacture of weapons, and for various technical needs of economic production. This was accompanied by a rising interest in investigating the depths of the earth, which was to produce a body of practical knowledge in prospecting for metals and other ore minerals. It was at this time that the first successes were achieved in studying specific structural forms, especially veined bodies, and their modes of occurrence. In the treatment of theoretical problems, however, geotectonics, which had little or no experience to draw upon, was still groping in the dark.

In the sixteenth and seventeenth centuries questions that are now considered part of geotectonics (transformations of the interior structure of the earth and of the relief of its surface) were usually treated in works with broad and indefinite philosophical content. Problems of geology were examined in such works along with metaphysical and

theosophical questions. Expressions of opinion on such problems alternated with discussions of morality, of "Providence," of the "fertility of the heavenly soil," of the reasons for the longevity of the first people, of the "terrible judgment to come," and so on. The first generalizations on tectonics in the Middle Ages and in modern times differed little from the fancies of the ancient Greek philosophers; they were no more free of fantasy, speculation, and the influence of the darkest aspects of Aristotelian and Scholastic philosophy. In the sixteenth century, for example, many people believed in the truth of Ristoro d'Arezzo's teachings (thirteenth century), which were based on the notion that the heavenly bodies exert an influence on the earth. The stars, in Arezzo's opinion, attract the land as a magnet attracts iron, so that the greatest dry land masses are formed where there is the greatest number of stars above the earth. Mountains have a different origin. When we look at the sky, we see that some stars are higher and others lower; thus the stars depict mountains and valleys in the sky. This celestial "relief" is imprinted on the earth exactly as a stamp impresses a figure on smooth sealing wax.

An idea of the nature and diversity of the theories explaining the origin of mountains that were in circulation to the end of the sixteenth century may be obtained from a book by Valerius Faventinus (1561)—the first work especially devoted to this problem. It appears that ten theories of the origin of mountains could be found in the works of the most authoritative philosophers.

The first theory connects the formation of mountains with earthquakes, a second with the swelling of the earth with moisture, a third with the lifting power of underground air. A fourth theory considers fire as the principal cause of the rise of mountains: subterranean fires cause the earth to boil like water, and the congealing, boiling earth's surface is transformed into mountains. According to a fifth theory, mountains are born, live, grow, and die like plants. A sixth theory turns to the stars, a seventh considers that mountains are formed through erosion of the earth by rivers, and an eighth associates the formation of mountains with the piling up of sand and earth by the wind. According to a ninth theory, vapors rising from the earth when it is heated by the sun carry with them particles of earth, which settle and accumulate to form mountains. The last theory, finally, says that some mountains are formed as a result of the work of man.

Most of the "theories" enumerated by Faventinus became historical curiosities and were soon forgotten.

Until the eighteenth century only a few investigators of genius could tear themselves loose from the shackles of fantasy and intellectual speculation. Among them was the great thinker and artist, Leonardo da Vinci (1452–1519). Agricola (1494–1555) and Nicholas Steno (1638–1687)

also deserve to be included in this group. These investigators differed from their contemporaries in basing their conclusions not on abstract reasoning and interpretations of the texts of ancient authors but on their own observations of natural phenomena, which were generally made with practical aims in mind. In them we may see how the growing practical knowledge of geology began to fertilize theoretical thought.

Leonardo da Vinci, who supervised the construction of canals in northern Italy, had the opportunity to make various geologic observations; from these he drew scientifically valuable conclusions that far outstrip his age. For example, he determined that rocks occur in the earth in layers. Although at that time the majority of scientists doubted that fossils were of organic origin and believed them to be "sports of nature," or chance combinations of matter, Leonardo da Vinci correctly interpreted the significance of fossils and, finding them in layers of rock that were then located on land, he concluded that shores do not remain fixed and that land and sea can exchange places.

In regard to the origin of mountains, Leonardo da Vinci held the extremely original view that mountain massifs are vestiges of the former level of the earth's surface and that all the remaining areas (lowlands and plains) have sunk down (Leonardo da Vinci, 1938).

In Agricola, who made observations in the mines of Saxony, we find the first correct ideas (except for some notes by Leonardo da Vinci on this question) of the destructive action of water currents gradually cutting down the valleys and thus increasing the relative excess of the interstream divides. Agricola observed the occurrence of ore bodies, chiefly veins, and produced the first correct concepts of the geometry of the intersection of veins with the uneven surface of the earth.

Nicholas Steno, a Dane by birth, lived and worked in Italy in the second half of the seventeenth century. In 1669 a small work of his was published, with the strange title *De solido inter solidum naturaliter contento* ("on the solid naturally contained within a solid"). This was probably the introduction to a more extensive composition, which, however, was never written. Steno's conclusions on problems of geology were based on the observations he made during excursions through the valleys and hills near Florence.

Steno determined that rocks from strata can be traced across valleys and through mountains. The stratification of the rocks led Steno to the conclusion that they were formed by settlement out of an aqueous medium. Through his observations of strata, their horizontal extension, and their modes of occurrence, he arrived at two principles: those of the extension of a bed and of its initial horizontal occurrence.

The first principle was expressed by Steno as follows:

At the time when any stratum was being formed, it was bounded by the sides of another solid body or else covered the whole earth. From this it

follows that wherever outcrops of strata are seen, one must either find their continuation or else discover another solid body which interrupted the material of these strata and prevented its continuation and expansion.

He formulated his second principle in this manner:

In regard to the shapes of the strata, there is no doubt that at the time a stratum was being formed, its bottom surface and its lateral surfaces matched those of the underlying and surrounding bodies; but its upper surface was everywhere parallel to the horizon, and consequently all the strata except the lowermost were enclosed between two planes parallel to the horizon. From this it follows that strata which are now perpendicular or at an angle to the horizon were parallel to it at an earlier time.

Disturbances in the original occurrence of the strata, as well as the development of the irregular relief of the earth's surface, Steno explained by partial collapse of the upper strata into hollows formed beneath them by the action of subterranean fires or erosion by water.

On the basis of his conceptions, Steno analyzed the geologic history of Tuscany. After observing the occurrence and interrelationships of the strata, he came to the conclusion that the history of Tuscany could be divided into six stages. During the first stage Tuscany, like the entire earth, was covered with water, out of which the primary beds without fossils settled and were deposited in horizontal layers. In the second stage the water retreated, the land became dry, and cavities were formed beneath its surface by the action of water and fire. In the third stage irregularities appeared in the earth's surface through the collapse of the strata above the hollows; in this process the strata assumed positions at various angles to

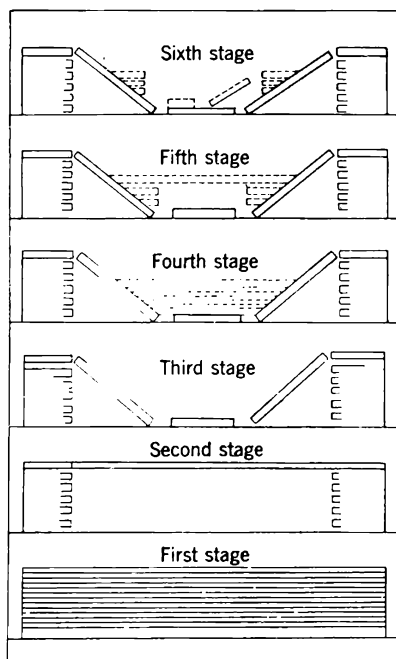


FIG. 2. The tectonic development of Tuscany (after Steno).

the horizon. The fourth stage was marked by a new advance of the sea onto the land. Water covered the whole region and deposited new horizontal strata, which this time contained fossils. These strata lay unconformably upon the older strata, whose position had been disturbed earlier. In the fifth stage the water again retreated, and the earth was heavily eroded by rivers and underground waters. Finally, in the sixth

stage, the present mountains and valleys were formed through a last collapse of the strata. His geologic history of Tuscany was illustrated by diagrams (Fig. 2).

Leibnitz (1646–1716) was also one of the great scientists of this era. He was among the first to express the idea that the earth was originally molten and then gradually congealed. The process of congealing and hardening was uneven, resulting in the formation of irregularities and bubbles in the solidified crust. It was in this manner that mountains appeared on the surface of this planet.

PLUTONISM AND NEPTUNISM

Until the middle of the nineteenth century there were two fundamental tendencies in geologic theory, which were later given the names *plutonism* and *neptunism*.

The proponents of the first theory, which was also known as vulcanism, saw the cause of all transformations on the earth's surface in the action of internal forces—fire in the interior of the earth. Those who adhered to the second theory denied the significance of interior forces and associated all geologic phenomena with processes occurring in an aqueous medium (precipitation, crystallization, erosion, and so forth).

Let us examine the history of the plutonic theory, which played a leading role in the geotectonic conceptions of the eighteenth and nineteenth centuries.

The plutonic theory has roots far back in time. Its origins must be sought in ideas about subterranean winds and vapors, which were mentioned by the ancient Greek philosophers.

At the end of the sixteenth century, the famous scholar Galesius, in his tabulation of the possible causes of earthquakes, indicated “underground exhalations” resulting from a combination of fire, earth, water, and air as the most likely: the underground air arising from this combination seeks an exit, causing the earth to shake.

Gradually, however, the conceptions of fire in the interior of the earth lost their mystical and fanciful coloring and came closer and closer to resembling something that might properly be called a scientific concept. An important contribution to this concept was made by observations not only of volcanos, which testified to the existence of “interior fires” in the earth, but to an even greater degree of ore deposits, which began to be worked successfully in the sixteenth and seventeenth centuries. The formation of ore-bearing veins within deep fissures in the earth's crust was naturally attributed to hot solutions and gases of deep-seated plutonic origin.

On the basis of these ideas of expanding subterranean gases, the first attempts were made, at the end of the seventeenth and beginning of the

eighteenth centuries, to work out a general theory of the development of this planet. The most interesting writings on this question are those of Robert Hooke (1635–1703). Hooke leaves no doubt of the fact that the earth's surface has undergone many changes. The Biblical Deluge, of which many then spoke, is not a sufficient explanation of all the signs of enormous transformations that have been discovered in the earth's strata. A hypothesis must be devised to explain the numerous mutual displacements of land and sea (Hooke, 1705; Pavlov, 1928). These changes have occurred

. . . by the Eruption of some kind of subterraneous Fires, or Earthquakes, whereby great quantities of Earth have then been raised above the former level of those Parts. . . . It seems not improbable, that the tops of the highest and most considerable Mountains in the World have been under Water, and that they themselves most probably seem to have been the Effects of some very great Earthquake . . . (Hooke, 1705, page 291).

According to Hooke, earthquakes can cause the earth's surface either to rise or to sink. Subsidence occurs when the roof of an underground cavity collapses, as the subterranean fires escape to the outside. Earthquakes may shake the entire surface of the earth, "overthrowing Mountains, and turning them upsidedown, throwing the parts of the Earth from one place to another, burying the superficial parts, and raising the Subterraneous" (Hooke, 1705, page 309). Here Hooke was clearly thinking of dislocations of beds, upwarping, crumpling, and fracturing.

A theory of the formation of mountains by the expansion of underground gases was very concretely stated by the Italian student Moro (1740). Underground "exhalations," heated by the planet's interior fires, expand. As a result, the earth's surface is uplifted, and cavities filled with gas or air develop beneath it. It is quite probable, however, that a void cannot exist within the depths of the earth. Hence the slowly forming cavity will gradually be filled, and the rise of the surface in one place will be accompanied by subsidence in another.

The proponents of the neptunian theory at first relied on facts that seemed then to be very well founded. They used the great achievements of physics and chemistry in the seventeenth century to solve problems of the crystallization of salts from aqueous solutions. This phenomenon was first studied in this period; its mysteriousness—the emergence of solid particles from a completely transparent liquid could not but seem mysterious to the unprepared observer—attracted the attention of scientists. Mineralogists surmised that this very crystallization might be closely related to the formation of the minerals in rocks. Although they were essentially following the right path, however, they committed great errors, which concentrated excessive interest on one process. Succeeding neptunists completely denied the existence of any internal

energy in the earth. Volcanos, in their view, were surface phenomena (caused by the underground burning of coal, oil, and sulfur), and all geologic processes depended on the action of water.

The adherents of neptunism supposed that all rocks without exception, including even those now known to be of igneous origin, were formed by the crystallization of various salts from water. Pebbles on the seashore and marine and fluvial sands were thought to be precipitations of salts. The neptunists did not acknowledge any large movements of the earth's crust; beds that were inclined or bent into folds they believed to be the result of defective crystallization on an irregular surface.

This notion of the role of water created a favorable atmosphere for the penetration of ideas of religious origin into geology. The Bible mentions the "world-wide Deluge," which covered the whole surface of the earth with water, except the peak of Mt. Ararat, for forty days and forty nights. Many neptunists, attempting to reconcile science with religion, believed not only that a world-wide flood actually occurred but that it was the greatest event in the earth's geologic history and that the majority of rocks had been formed from the waters of this flood. Thus arose the theory in earth science known as *diluvianism* (from the Latin word *diluvium*, meaning "flood"). This was particularly popular in England. Its adherents produced many fantastic "theories of the earth," in which scientific concepts were drowned in a mass of fictions. The neptunist and diluvianist Woodward (1665-1722), for example, wrote that the ancient Greek philosophers believed in changes in the earth's face because they were pagans. In actual fact, there had never been any cataclysm besides the one described in the Bible: the "world-wide Deluge." Another writer believed that mountains were formed by the crystallization of salts precipitated from the waters of the ancient world-wide ocean around excrescences of the sea bottom, and he drew an analogy between this process and the deposition of crystals on a stick lowered into a supersaturated solution.

The neptunian doctrine reached its zenith in the works of the famous Saxon mineralogist Werner (1749-1817). Werner was a brilliant lecturer and enjoyed enormous popularity during his life. The school of mines at Freiberg, where he was a professor, attracted many students, so that an army of followers surrounded Werner. After his death, however, neptunism, having been carried to its absurd logical extreme, proved inadequate. Many of Werner's followers, such as von Buch, Humboldt, and other great scientists, went over to the plutonists, and the Wernerites disappeared from the scene at the beginning of the nineteenth century. During the eighteenth century, however, the two theories continued to compete.

The eighteenth century saw the growth of scientific geology in Russia. Russian geology, as will be seen below, developed largely independ-

ently. A Russian school of geology was formed very early; its nature was clearly reflected in the brilliant writings of M. V. Lomonosov (1711–1765) and later in works by Russian scientists of succeeding generations.

The rapid development of scientific geology in Russia resulted from the fact that it was based on the practical experience of old Russian mining experts, salt plant workers, master workmen in quarrying building stone, and other "industrial persons" who had been extracting rocks, minerals, and metals for a number of centuries in various regions of the country.

In the eighteenth century the existence of the two theories—neptunism and plutonism—was also reflected in the development of geology in Russia. The chief representative of the neptunian theory was Pallas, who came from abroad and spent almost the whole of his life working in Russia. As the leader of a great expedition of naturalists in Russia, he made the first geologic (more accurately, geognostic) reports on many provinces of the country, both European and Asiatic. In contrast to other participants in expeditions during the second half of the eighteenth century, which were limited almost entirely to description, he also proposed some theoretical generalizations (Khabakov, 1950).

It is clear from these generalizations that Pallas held neptunian views but was not as narrow a neptunist as Werner; in some cases he expressed doubt of the absolute truth of neptunism. In one particular work Pallas considered the origin and structure of mountains. He wrote that conclusions in regard to the regularities of their structure could be drawn only on the basis of a comparison of many mountain ranges; such comparison shows that the cores of high mountains are composed of granite. The granite is covered by schists that are steeply inclined to the horizon. These constitute the oldest, or primary, formation, which contains no fossils. These are followed by the symmetrical limestone belts of the secondary formation; they are inclined at a smaller angle to the horizon, and in their upper strata they contain abundant fossil fauna. Still further from the axis of the range occur the tertiary, unconsolidated rocks, represented by sands, clays, and marls, which also contain a great quantity of fossil remains. Typical examples of such structure are the Urals, where, according to Pallas, this very succession of rock occurrences is observed. Among the tertiary deposits he included the sand and clay deposits on the western slopes, which belong, as we now know, to the Permian system.

Pallas described the formation of mountains in the following manner: First, islands appeared in the ancient ocean, on the bottom of which the primary granites had been deposited. Disintegration of the granites produced the material for the formation of the schists that were deposited on the underwater slopes of the granite islands. Later

the secondary limestone settled out from the water and finally the sand and clay rocks of the tertiary formation. At the time these rocks were deposited, great masses of pyrite accumulated in places and ignited on contact with the organic residues and water. This was the origin of volcanos. The explosive force of the expanding gases caused the fracturing and raising of the earth's surface and also the uplifting of the mountain ranges and the steep inclination of the strata. Pallas departed somewhat from the consistent views of neptunism. Although he regarded volcanos from the neptunian point of view—that is, he saw them as surface phenomena caused by the ignition of burning substances within the sedimentary mantle—he nevertheless attached far greater importance to them than Werner, who almost completely denied that there was any kind of movement of the earth's crust.

This view of the structure of mountain ranges as a kind of inverted fan, along with the succession of formations on mountain slopes suggested by Pallas, circulated widely and was accepted for a long time.

Beside Pallas, the geologists in the neptunian group at the end of the eighteenth and beginning of the nineteenth centuries included Zuyev, Lepekhin, and Severgin.

An original feature in Russian geology of that time was the creation of a rational synthesis of what was useful and correct in both the neptunian and the plutonian views. This development was produced by M. V. Lomonosov, who based his thinking on the fundamental importance of internal forces in the development of the earth but nevertheless took certain ideas of surface processes from the neptunists and reworked them into his original system. Lomonosov is properly considered the author of the first scientific hypothesis in geotectonics (the uplift hypothesis) and the founder of all modern scientific geotectonics.

THE HYPOTHESIS OF UPLIFT

Lomonosov set forth his teachings mainly in his famous work *On the Strata of the Earth* (modern edition published by Gosgeolizdat, 1949).

After considering the forces that produce and change the face of the earth, Lomonosov concluded that these forces are both external and internal. "The external forces are strong winds, rains, river currents, waves, ice on the sea, forest fires and floods; earthquakes alone are internal" (page 45). Earthquakes cause the elevation and sinking of sections of the earth's surface, the displacement of shore lines, and the appearance and disappearance of mountains, islands, and whole continents.

"... There is in the heart of the earth ... an immeasurable power which at times makes itself felt on the surface and whose effects are visible everywhere that we see the ocean bottom on mountains and mountains on the bottom of the ocean. . . ." (page 49).

On the subject of the cause of earthquakes, he wrote: "The force which lifts such a load cannot be ascribed . . . to anything like fire reigning in the earth's womb. . . ." (*ibid.*).

M. V. Lomonosov explained disturbances in the occurrence of beds by vertical buckling:

The position of rocks inclined at a very great angle to the horizon shows that these strata were dislodged from their former position, which by the laws of mechanics and hydrostatics ought to be horizontal. . . . Thus when mountains rose from the sea bottom they were impelled by an internal force; invariably the rocks composing them were forced to buckle, to split apart and form clefts, tilted positions, river rapids, precipices of various heights and towering shapes (pages 56-57).

Lomonosov painted an even more colorful picture of the change in the occurrence of strata under the influence of internal fires:

The effects of various forces have but one cause; this cause is the action of fire, although the surface rises in one place and sinks in another. In the first case, the superabundant hot substance, creating a widespread and intense flame, tears at the surface lying above it and through the fissures seeks a path to the air. Then, having escaped into the open, the hot substance allows the pieces it has torn out from the surface to fall back by the pull of their own weight, but this cannot restore the shattered fragments to their former position and order. They fall, like a collapsed brick arch, one upon the other, on their edges, crosswise, on their ends and on their sides; and in such a collapse they occupy a much greater space, leaving frequent empty gaps between them: thus it is that mountains rise above the rest of the earth's surface. At other times a great quantity of burning matter has blazed for long ages, spreading smoke and flame, and has thus burned out a great hollow, over which the surface, suspended by its edges, could no longer cling to the surrounding earth; then, collapsing under its own weight, it has fallen and provided a space for the waters to form a sea or a lake (page 64).

From the foregoing quotations it is clear that Lomonosov regarded certain movements of the earth's crust, particularly dislocations of strata, as occurring rapidly. In addition, however, he evolved a completely new theory of slow, unnoticeable movements of the earth's surface, its "imperceptible and long-lasting rise and fall," which cause the advances and retreats of the seas.

M. V. Lomonosov was thus the first to classify tectonic movements as fast and very slow, as is done today in distinguishing relatively rapid folding and faulting, on the one hand, from the very slow vertical oscillations of the earth's crust, on the other.

In speaking of slow, "imperceptible" movements of the earth's crust, Lomonosov was proposing an evolutionary theory of the earth's development; he differed sharply with the ideas of the catastrophists, who at that time flourished in other countries.

The catastrophists maintained that the past history of the earth was punctuated by overwhelming catastrophic events caused by extraordinary, supernatural forces. During such catastrophes all life on earth was destroyed, new mountain chains arose, former mountains sank to the bottom of the sea, and cataclysms took place in the depths of the earth. When the catastrophe ended, the supernatural forces subsided and remained at rest for a time, until the next catastrophe. The doctrine of catastrophism was founded on two circumstances: first, that observations had shown that the structure and relief of the earth's surface had undergone repeated changes, and second, that the catastrophists were convinced of the extremely brief duration of geologic time and of the small age of the earth.

Scholars first determined the age of the earth from the Bible—that is, as little more than six thousand years. Buffon, risking the wrath of the ecclesiastical authorities, dared to increase this figure to seventy thousand years. Thus, in the conceptions of eighteenth-century students, the total length of geologic time did not exceed several score millennia. The entire complex succession of geologic events that have left their record in the crust of the earth had to be fitted into this brief segment of time. The difficulty could not be circumvented without catastrophes. It was acknowledged that the processes that are seen to take place on the earth's surface at present could not have had time to bring about perceptible changes within this short span of the earth's existence. Running water erodes the earth, but how much time will it require to cut a deep valley? Believing that nature had not had time enough to carve valleys by this slow method, scientists turned to other processes: they supposed that during the catastrophes the earth was covered with fissures and that these fissures became valleys. As late as the first half of the nineteenth century, even such great geologists as von Buch did not appreciate the importance of erosion by water and saw valleys and gorges as cracks in the earth's crust.

Lomonosov's theoretical syntheses were not only generalizations from what was, for that time, a great amount of factual material; they were intended to create a theoretical basis for the solution of practical problems. It is known that Lomonosov studied numerous deposits of metal ores in Central Europe in detail. He carried on an extensive correspondence with mining experts in the Urals and other districts of Russia, using the information he received to create a theory that would explain the formation of ores and their emplacement in the depths. It was not without reason that he named one of his works *The Generation of Metals from the Earth's Quaking*. Within the limitations of the science of his day, Lomonosov found the best solution to the problem he had posed himself by associating the formation of metals in the earth's crust with underground fires and hot gases in the depths.

The geotectonic hypothesis of uplift was further developed by Hutton (1726–1797). In 1788 he published his *Theory of the Earth*, which the majority of his contemporaries greeted with hostility. Its appearance, however, was of great importance for the development of geologic science as it is known at present (Belousov, 1938b).

The *Theory of the Earth* defined the tasks of geology, its capabilities, and its methods. Hutton believed that geology should not concern itself with the origin of all things and other such problems of philosophy. It is unworthy of a scientist to cover up his ignorance by recourse to supernatural catastrophic forces. All geologic phenomena must be explained by the action of natural agents alone—agents whose operation can be observed at the present time. The earth, in Hutton's opinion, exists and develops in conformity with the ordinary laws of physics and mechanics; and he was fond of repeating that the earth is a machine.

All geologic processes, in Hutton's view, are interrelated and occur in a definite succession. In his work we find rocks classified as sedimentary and igneous; he was the first to discover the phenomenon of contact metamorphism around masses of granite.

Among the various movements of the earth's crust, according to Hutton, the most important are vertical movements, meaning precisely those uplifts caused by the lifting force of the planet's internal fires. He outlined the history of the earth in the following manner. A certain part of the earth's surface is elevated and a continent is formed. The latter is destroyed by the action of external forces; the products of this destruction are accumulated as clastic rocks on the bottom of the surrounding marine basins. Finally, the continent is completely destroyed. A new uplift takes place on the site of the former seas, and from these seas a new continent arises which is composed of the fragments of its predecessor. Such alternating movements have occurred throughout the history of the earth. During the vertical uplift of the earth's crust, dislocation of the primary strata there also occurs.

Though the primary direction of the force which thus elevated them [the strata] must have been from below upwards, yet it has been so combined with the gravity and resistance of the mass to which it was applied, as to create a lateral and oblique thrust, and to produce those contortions of the strata, which, when on the great scale, are among the most striking and instructive phenomena of geology (Playfair, 1802, page 45).

In regard to the duration of geologic history, Hutton declared himself opposed to catastrophism and an active proponent of evolutionary ideas. Believing that in the past only natural forces have acted on the earth's surface and in its depths, in obedience to the laws of physics and mechanics, he was led to the inevitable conclusion that only those processes have occurred in the past of this planet which are also occurring today.

Hutton struggled to extend the limits of geologic time, believing that it could be considered, roughly speaking, as infinite: in the economy of nature "we find no vestige of a beginning,—no prospect of an end," he said.

At the end of the eighteenth century a quarrel raged between the adherents of the doctrine of catastrophism and those who believed in evolutionary ideas. In this competition the advantage was first clearly on the side of the catastrophists. Hutton's ideas had no success during his lifetime; he was accused of forsaking the faith and of crimes against religion. Among his opponents was by far the most authoritative naturalist of that time—L. Cuvier.

It was not until the middle of the nineteenth century that Charles Lyell (1797–1875) succeeded in closing the issue opened by Hutton and, having achieved complete victory, in creating a new scientific geology based entirely on the principles of evolution. In his *Principles of Geology*, which first appeared in 1830, he examined the reasons for the general stagnation in geology. He concluded that the principal cause was the concept of the brevity of geologic history (Lyell, 1866). Analyzing numerous examples in detail, he showed that the structure of the earth's crust contains no indications whatever of the action of supernatural forces and that all events of the geologic past can easily be explained by ordinary natural forces, which are acting even today.

The line of reasoning along which Lyell led his reader may be traced briefly. It cannot be denied that rivers, waves, and volcanos change the earth's surface or that the forces that produce earthquakes are capable of raising or lowering the surface. One can reasonably believe, however, only that these changes in the earth take place not by successive catastrophes, but gradually.

To show this, Lyell dwelt at length on the history of the Tertiary period. This choice was not arbitrary: many of Lyell's contemporaries, led by Cuvier, having studied the fossils of the Tertiary system in detail and having discovered repeated changes in the fauna during this period, came to the conclusion that these changes served as an excellent proof of a series of catastrophes leading to the total destruction of one fauna and the generation of a new one. Lyell succeeded in showing that, although the Tertiary period may be divided into four epochs, these epochs do not differ sharply from each other, since there were many species of organisms that lived through several epochs, and that many forms that had lived in Tertiary time continue to exist even at the present time. The last uplift of the earth's surface, which drained a considerable part of the present European continent, was so tranquil that shell species found on the island of Sicily, for example, which were discovered in the fossil state at a height of more than 1,000 m above sea level, still live in the waters of the Mediterranean Sea.

Comparing Tertiary with recent rocks, Lyell found a great similarity between them and proved that Tertiary rocks were created under the same conditions as those under which the recent rocks have been forming in the present. Among ancient deposits he found some that to all appearances were deposited by the wind and have the same composition and structure as recent dune sands. He noted the complete similarity between ancient and recent lava flows, compared tectonic fissures with those which appear during earthquakes in our time, described fossil seashores, and compared ancient calcareous tufas with the present-day deposits of mineral springs. Turning to earlier formations, he discovered remains of fossil soil among Upper Jurassic deposits and traced bays, sand bars, and swamps in rocks of Carboniferous age. He also found current-made and wave-made ripple marks on the surfaces of ancient rocks.

All these and other phenomena, brilliantly studied and described, were well calculated to persuade the reader that there is no difference whatever between recent and past geologic processes and that the structure of the earth's crust reveals no traces of "supernatural" catastrophes. Lyell's principal conclusions were that the length of geologic history is exceedingly great and that ancient geologic processes can be interpreted entirely by the laws that govern the processes now changing the face of the earth.

It is well known that Lyell had a great influence on Darwin, who called Lyell his teacher. Darwin, in turn, both in his biologic works and in a few special geologic treatises, contributed greatly to the establishment of evolutionary ideas in geology.

Let us return, however, to the geotectonic uplift hypothesis. This was formulated, after Hutton, during the first half of the nineteenth century in the works of von Buch (1774-1852) and then in studies by Swiss geologists (Studer et al.). During the whole first half of the last century, this was the reigning hypothesis in geotectonics (Belousov, 1939). The adherents of the uplift hypothesis in Russia were D. Sokolov, E. Eykhval'd, G. D. Romanovskiy, N. A. Golovkinskiy, G. Ye. Shchurovskiy, and other researchers.

In studying the structure of volcanic cones, von Buch discovered that they were formed of rocks in strata that always sloped from the center of the crater toward the periphery. Not realizing that this inclination of the strata was due to the gradual piling up of the products of volcanic eruptions, von Buch decided that the domical structure of volcanic cones resulted from the lifting force of the magma rising from the depths through the throat of the volcano. The structure of such "craters of uplift" strongly resembled the pattern generally observed in mountain ranges: inclination of the strata from the center of the range toward the periphery. From this analogy there emerged an original conception

of the formation of mountains, which occupies a central position not only in von Buch's theoretical system but in all geotectonic hypotheses of the first half of the nineteenth century. Von Buch believed that all the manifestations of uplift, dislocation, and crumpling of strata that are observed near the surface of the earth were due to the direct effect of the volcanic rocks intruded into them from below. A number of chance observations and, from the standpoint of the modern investigator, surprising inferences led von Buch to conclude that the dislocation of strata in mountain areas is caused by the intrusion, not of volcanic rocks in general, but only of very specific rocks—augite porphyries. Von Buch illustrates his inductions with cross sections in which one may see mountain ranges that seem to float on the surface of dark porphyries being intruded upward from below (Fig. 3).

Von Buch's view of the role of porphyries was refuted, however, and the uplift hypothesis began to reappear in a broader form: the cause of the uplift and dislocation of strata began to be seen in the intrusion of

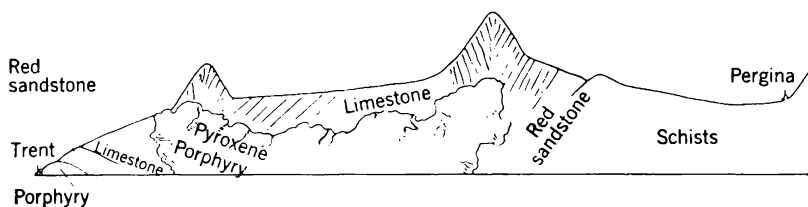


FIG. 3. Elevation of mountain massifs by dark porphyries intruded from below. Section from Trent to Pergina (*after von Buch*).

various crystalline rocks and not of porphyry alone. Since enormous granite massifs were known to exist in the central parts of the Alps, the study of which formed the basis of all the geotectonic views of the last century, granites, along with other igneous rocks, began to be considered the most probable cause of dislocation and uplift.

The uplift hypothesis was elaborated in this form chiefly by Swiss geologists, among whom Studer (1794–1887) deserves special mention. From his basic works, published at the end of the eighteen forties and beginning of the eighteen fifties, we find that the adherents of the uplift hypothesis, at the time of its greatest popularity, explained the dislocation and uplift of strata as follows (Studer, 1847): magma expands and rises from the interior of the earth and is intruded into the sedimentary strata that compose the outer crust of the earth, upturning the rocks it encounters in the course of its rise. The disrupted zone acquires the following structure: in the middle is the central crystalline massif formed by the rising magma, surrounded on the periphery by sedimentary strata sloping downward from the central massif to its outer edge.

The effect of the magma is not confined to uplifting. Upon being intruded, it pushes the overlying rocks to the side, producing a horizontal movement in the latter that also results in the formation of folds (Fig. 4).

Thus the adherents of the theory of uplift assumed that the primary forces in the earth's crust are vertical. If horizontal stresses also arise in the crust, crumpling the strata into folds, these will be due to transformation of the vertical stresses into a horizontal thrust. The primary importance of the vertical forces is the fundamental principle of the leading geotectonic hypotheses of the first half of the nineteenth century.

Studer's works are the culmination of the development of the uplift hypothesis, which, as we have seen, has a very long history. By the end of the eighteen fifties, however, this hypothesis began to lose ground, and by the beginning of the eighteen seventies had left the scene completely. The reasons for this, and the hypothesis that replaced it, will

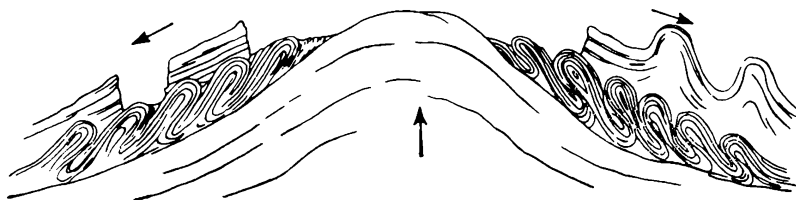


FIG. 4. The formation of folds (*after Studer*).

be discussed below; at this point it will be appropriate to mention some of the achievements in structural and regional geology in the eighteenth and beginning of the nineteenth centuries.

ACHIEVEMENTS IN STRUCTURAL GEOLOGY AT THE END OF THE EIGHTEENTH AND BEGINNING OF THE NINETEENTH CENTURIES

During the eighteenth century, important advances were made in studying the morphology of structures. At first solely in connection with deposits of ore minerals, and later as the result of extensive regional explorations in prospecting for ore minerals, detailed descriptions were made of the structure of many regions and the first maps and sections showing the distribution of rocks appeared.

As early as the beginning of the eighteenth century, the Englishman Strachey (1671–1743) described the geologic section through the coal measures of Somersetshire. His description is distinguished for its great detail: it includes an enumeration of the successive strata, with identi-

fications of the rocks and precise calculations of their thicknesses. Strachey stated that all the strata have an inclination of "approximately 22 inches per sagene" [2.134 m]. The first drawings from nature of the complex structures of the Alps (Fig. 5) were also made at the beginning of the eighteenth century.

In the middle of the eighteenth century Lehmann and Fuchsee

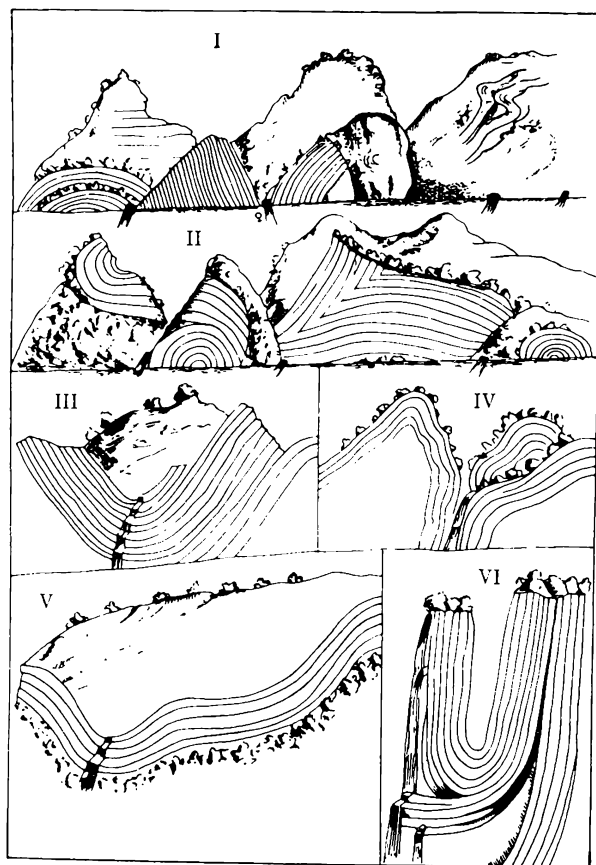


FIG. 5. Geologic sections through districts in the Alps, drawn by Valigneri, in 1718.

studied the geologic structure of Saxony and Thuringia in detail and made the first geognostic maps and sections of these districts.

Explorations of regional geology became particularly extensive at the end of the eighteenth century. The last three decades were the period in which the first real facts in the science of the earth were collected, in which the solution of problems in the history and structure of this planet was finally removed from the studios to nature, to the field, and

to outcrops of rocks, and in which the hammer became the chief tool of geology.

During this time an entire galaxy of eminent scientists studied the structure of Russia, particularly those parts of it, such as the Urals, which were rich in ore minerals. Pallas (1741–1811) was among the first geologic explorers of Russia. Then Hutton made his observations in the mountains of Scotland; Guettard, Giraud, Soulavie, and Desmarest studied the geologic structure of France; Saussure (1740–1799) made his investigations of the Alps (Khabakov, 1950; Zittel, 1899; Tikhomirov and Khain, 1956).

At the turn of the eighteenth and nineteenth centuries Cuvier began his investigations in the vicinity of Paris. The first regional explorations were begun in North America. The works of the English irrigation engineer William Smith (1769–1839) date from the first decade of the nineteenth century; these works, which were of a practical nature, led to the discovery of the stratigraphic significance of fossils and the creation of stratigraphic paleontology, which to this day is the foundation of research methods in geology. Smith's discovery made possible the transition from geognostic maps, which showed the lithology of the rocks, to geologic maps, on which the rocks could be classified according to age. This allowed a considerably more thorough understanding of tectonic structure, although later it caused some loss of interest in the material composition of rocks.

The result of all these investigations and discoveries was a great increase in knowledge of the geometry of dislocations. It was understood that the principal structures in mountain areas were folds, and careful study of the distribution of folds was for a long time almost the chief task of regional geologic explorations.

Regional exploration and the accumulation of factual data on the geometry of the structures in the earth's crust led investigators to reject the uplift hypothesis. The most serious blow to this hypothesis followed the establishment of the geometric regularity of folded structures.

Study of folded regions showed that they do not contain separate, disconnected folds forming parts of the central massif, but belts of long regular folds extending parallel and en echelon to each other. Folded zones, as we have seen, extend across entire continents. The uplift hypothesis was unable to explain these characteristics of folded zones. If a volcanic cone is the model for a dislocated area, this area should most likely have circular contours. The distribution of zones of dislocated strata in belts was still not understood. Moreover, in attributing the active role in the formation of folds to the central granite massifs of the Alps, the adherents of the uplift hypothesis thought that these massifs were young granite intrusives injected into the surrounding rocks after the latter had been formed. The development of the methods of geologic

research soon made it possible to prove that the granites of the central Alpine massifs were not younger, but older than all the surrounding rocks. As it turned out, the granites, along with the other rocks, played only a passive part in the folding. This fact, established in the eighties by Heim and Suess, finally forced investigators to abandon the uplift hypothesis.

THE CONTRACTION HYPOTHESIS

After the uplift hypothesis had been rejected, it was necessary to create a new thesis in geotectonics; its nature was predetermined by the success of the study of folded structures, as discussed above. This success was, in turn, determined by the nature of the practical work being done in geology. Until the beginning of the nineteenth century, the most extensively mined raw materials were metallic minerals. Metals are associated chiefly with massive rocks of igneous origin. For this reason, as an explanation of the structure of the regions with which geologists were then concerned, the most suitable appeared to be the uplift hypothesis, in which the leading role was attributed to volcanic forces and volcanic rocks. Sedimentary, stratified rocks attracted relatively little attention during this period. The situation was changed at the very end of the eighteenth and beginning of the nineteenth century, when mined coal acquired its great practical importance. Since this is a raw material found in beds, it became necessary to study the structure of stratified rock series in order to understand its distribution within the earth's crust.

Through such studies, which were made primarily in coal basins, it was established that folds are the most common form of dislocation of bedded rocks. Thus there came to be a rising interest in the study of folded structures and inquiries into its causes.

The most natural-appearing explanation for the wavelike bending of strata into folds was a horizontal contraction of the earth's crust. A new tectonic system therefore arose, according to which not vertical, but horizontal forces now predominated in the earth's crust and caused all the various tectonic phenomena. This was the hypothesis of contraction, or shrinkage, of the earth's sphere, which was elaborated in scientific form by the French geologist Élie de Beaumont (1798–1874) and published by him in 1852. In the second half of the nineteenth century this hypothesis almost completely dominated geologic thought.

The contraction hypothesis and the tectonic ideas connected with it require examination in greater detail. It will be recalled that, according to this hypothesis, the earth is in a state of cooling and contraction. The outer shell has cooled as much as possible and thus retains its size; the

interior of the earth, which is still insufficiently cooled, continues to contract and decrease in volume. As a result, the earth's crust becomes too roomy for the enclosed core and is forced to "settle" gradually, to decrease its surface area, thus crumpling into folds.

This hypothesis thus supposes that there are permanent general forces of compression in the crust that envelop the whole of it uniformly. This is the central point of the contraction hypothesis. Geologic explorations, however, have shown that tectonic processes follow different courses in different places; in any one period, some parts of the crust undergo folding, while others maintain their original structure intact. To explain this circumstance the idea was introduced into the contraction hypothesis that the earth's crust is divided into rigid and plastic sections, which react differently to the same mechanical stresses. According to this concept, the rigid portions of the earth's crust (fault blocks, mountain massifs, platforms) act as vises, compressing and crushing the plastic zones in the total contraction of the earth. The folds are pressed against the rigid portions of the crust like waves hurled against cliffs and the contours of the folded zones and the shapes and disposition of the folds depend on the shapes, contours, and disposition of the rigid masses. The idea that the earth's crust was heterogeneous was first expressed by Suess (1831–1914) in his work, *Origin of the Alps* (1875). Later, proceeding from this concept, he summarized the structure of the earth as a whole on the basis of the contraction hypothesis (Suess, 1883–1909).

The contraction hypothesis gained increasing acceptance, as a result of the fact that at the very end of the nineteenth century structures were discovered in folded zones that were mistakenly taken to be very large horizontal displacements of whole sections of the earth's crust, accompanied by the formation of such structural forms as decken or overthrusts. The structures of some overthrusts might seem to suggest horizontal sliding of large complexes of rock for a distance of sometimes more than 100 km. Since in the Alps, for example, such overthrusts are fairly numerous, the contractionists were apparently right in regarding them as proof that there had been a great contraction of the earth's original surface. The first concept of the overthrust structure of the Alps was formed by Bertrand (1847–1907).

Overthrusts were the subject of heated controversy in the eighteenth nineties; at the geologic congress held in Vienna in 1903 the theory of overthrusts was recognized officially.

Considerable advances in geotectonics occurred during the predominance of the contraction hypothesis, which provided a general foundation for understanding geotectonic processes. In the seventies appeared Heim's famous work on the mechanism of mountain formation, which

examined the mechanics of folding from the standpoint of the contraction hypothesis (1878). Somewhat later Bertrand, by analyzing angular unconformities, established the periodic nature of large-scale tectonic movements and distinguished folded zones of different ages. He was the first to classify folded zones as Precambrian (Huronian), Caledonian, Hercynian, and Alpine (Bertrand, 1886–1887).

The concept of geosynclines also developed in the climate of the contraction hypothesis. Hall (1811–1898) first called attention to the fact that in folded regions the thickness of the deposits is greater than in places where the strata are undisturbed. He suggested that near the edge of the continent, where much clastic material is accumulated, the earth's crust is bent downward under the weight of this material, so that great thicknesses of sediments can accumulate here. These are the areas that turn out to have the greatest tendency to be crumpled into folds (Hall, 1859).

Dana (1813–1895) developed the more definitive form of the geosynclinal theory during the seventies of the last century. He regarded geosynclines as great depressions of the earth's crust formed by the action of horizontal forces of compression. The depressions fill with sediments, and further compression crumples these deposits into folds. Dana believed that geosynclines are formed along the edges of continents (as the Appalachian and Pacific Ocean geosynclines are situated in North America) because the ocean bottom, as it sinks during the contraction of the earth's crust and becomes wedged between the continents, exerts a horizontal pressure on them (Dana, 1873).

The theory of geosynclines was further developed in the works of Haug and Argand (Haug, 1900; Argand, 1916) and most recently in papers by Soviet investigators.

A geotectonic hypothesis based on the predominance of horizontal forces was developed in a coherent form by Stille (1924). It appeared very late, however, at a time when this hypothesis had already begun to lose its importance. In Stille's view the vertical oscillatory movements of the earth's crust were viewed as an irregular warping of the latter by horizontal forces—as folding of a sort, though weakened and manifested in rigid, slightly deformed sections of the earth's crust. The same horizontal forces determine the formation of faults and are also the cause of igneous activity, causing the magma to rise by squeezing it upward.

Thus, whereas the adherents of the uplift hypothesis explained all tectonic movements by the action of vertical forces, investigators who accepted the contraction hypothesis tried to do the same with the predominance of horizontal stresses. Both were faced with the same problem in principle—to subordinate diverse tectonic movements to a single leading process—but their attempts each had a different basis.

METHODOLOGIC DEFECTS OF THE CONTRACTION HYPOTHESIS

The concept of gradual shrinkage of the earth, which forms the cornerstone of the contraction hypothesis, is generally correct in pointing out that the earth has undergone a very slow evolution.

It is true that we cannot, at the present time, accept the idea of contraction in the form in which it was expressed in the nineteenth century, before the discovery of radioactive decay. It was then thought that the earth was shrinking and cooling because of the loss of heat it had received from the sun. Now it is known that the earth has its own source of heat in the radioactive elements that decay at a constant rate and are present, dispersed throughout the planet, in all rocks. This is not, however, a reason for completely rejecting the idea that the earth must ultimately become cool, inasmuch as the amount of radioactive elements is gradually decreasing as a result of decay. According to the theory of the earth's origin developed by O. Yu. Shmidt, the earth was cold in the beginning and was later warmed by the accumulated radioactive heat. Shmidt assumed that this heating of the earth is continuing at the present time; his assumption was confirmed by the work of his students (Lyubimova, 1958). But even in this theory, the earth must begin to cool sometime in the future, because of the decrease in the total quantity of radioactive elements.

Although the idea that the general trend of development of the earth is in the direction of cooling still maintains its importance to this day, the concrete application of this idea in explaining specific geotectonic processes is another matter. From the standpoint of geology, the contraction hypothesis is usually taken not so much as a general theory of cooling and shrinkage of the earth's volume but rather as one of a specific mechanism of the formation of folding and other dislocations. This mechanism is horizontal compression of the earth's crust. Geologists, in speaking of the contraction hypothesis, ordinarily have in mind an explanation of folding and other tectonic movements by the lateral compression of plastic portions of the earth's crust between more rigid blocks; in other words, the contraction hypothesis of folding. It is in this sense that the hypothesis will be further discussed below.

In the first decade of its existence the contraction hypothesis, in this understanding, undoubtedly played a positive role in geotectonics, since it provided a theoretical basis for explaining folded structures. It gradually lost its value as time went on, however, and by the beginning of the present century it clearly began to retard the development of geologic science, since it diverted geologists from studying the material composition of rocks to the creation of purely abstract geometric systems.

The adherents of this hypothesis assigned the leading role in the life of the earth's crust to movements producing folding, and they almost

completely ignored the oscillatory movements of the crust. Nevertheless, it is oscillatory movements that determine the material composition of the rocks in the earth's sedimentary mantle, whereas folding mainly affects the mode of occurrence of these rocks. Its effect, in other words, is purely geometric. Since it did not give oscillatory movements their rightful place in the tectonic process and since it failed to explain correctly the exceptionally delicate and complicated mechanism involved, the contraction hypothesis naturally could not give geology a proper basis in methodology for the study of these movements, which remained outside the sphere of interest of investigators. On the other hand, extraordinary powers of ingenuity and imagination were lavished on the reconstruction of folded structures and the geometric modes of occurrence of rocks. Geologists showed great virtuosity in drawing lines of demarcation on geologic sections through folded regions. Gradually the drawing of lines and structures was transformed from a means of understanding the laws governing the development of the earth's crust into an end in itself. In this manner the contraction hypothesis contributed to a tendency toward formalism in geology.

This formalistic tendency in geotectonics manifests itself in the exceptional thoroughness with which investigators depict the structural forms of the rocks, outlining these forms with great skill on sections, maps, and diagrams, whereas their understanding of the structural geometry does not go beyond the limits of these primitive, purely mechanical representations. A consideration of the history and conditions of the formation and accumulation of rocks in their observed interrelationships or a study of the physics of their deformation forms no part of the program of research. Their study is restricted to the construction of abstract schemes. The proponents of this tendency believe that one can explain the observed structures by allowing such and such a distribution of compressive or tensile stresses, such and such disposition of the rigid and plastic portions of the earth's crust, and so forth. Moreover, they fail to ask themselves whether the assumed mechanism actually exists, what were the historical premises of its genesis, and how it is related to processes occurring in adjoining regions, etc.

The formalized approach to the study of tectonic phenomena quickly led to a situation in which even a simplified mechanical explanation was offered *pro forma*. Structural laws became completely geometric, as if geologists were dealing not with rocks whose distribution reflects the complex history of various tectonic movements but with some sort of ideal, imaginary geometric bodies subject to simple mathematical analysis.

Under these circumstances, even generalizations in the field of geotectonics followed the same purely geometric paths. An example of such a formalistic generalization is the above-mentioned theory of the

nappe structure of the Alps, since it focused upon itself all the features of this tendency. With great skill and striking boldness the adherents of this formalistic theory (Kober, Staub, and others) depicted elaborate and complicated nappe structures, nine-tenths of which were the fruit of purely geometric speculation not based on facts and completely inexplicable from the standpoint of the mechanism and history of the structure.

The damage done by the theory of the nappe structure of the Alps was exceedingly great. Geologists have for a long time considered the Alps to be a classic region, an example of structure that must be studied to understand the structures of other folded regions. It is understandable, therefore, that, although overthrust theory was fully applied only in the Alps, it greatly influenced the thinking of geologists throughout the world. The theory of overthrusts, which introduced so much that was inexplicable from the standpoint of mechanics and accidental in regard to the relationship between the various movements, excluded a number of the most important problems from the field of scientific analysis. The suppositions that all the rocks in a folded region do not occur on the site of their primary formation, that they have all undergone varying but large horizontal displacements, and that it is now for all practical purposes impossible to determine their original disposition make it impossible to study the history of folded regions. One must above all know the original distribution of the facies and thicknesses of the deposits. Inasmuch as this theory assumes that, because of the breaking up of the rocks into groups that undergo most varied horizontal displacements, individual groups of beds are reshuffled in completely random fashion, it deflects the geologist's attention from problems of stratigraphy, which are then not solved as critically as they should be.

It seems that the contraction hypothesis, after several decades of general acceptance, was dropped quickly precisely because of these formalistic digressions. The objections against this hypothesis, which were raised as early as the beginning of this century and became especially strong in the nineteen twenties, need not be examined in great detail.

The chief difficulty arose in explaining oscillatory movements. It has already been said that investigators tried to see them as a weak form of folding, reflected in folds of very great radius. But this is contradicted by the basic feature of oscillatory movements—the replacement of uplift by subsidence, and vice versa. An increasing compression can only intensify the bending of the earth's crust in the same direction, but it cannot change a convex bending into a concavity and back again. Moreover, a study of oscillatory movements reveals their surprising complexity and fineness. The earth's crust is divided into a multitude of comparatively small areas distinguished by the regime of their oscillatory move-

ments. These sections are distributed regularly, and the succession observed in the course of the oscillatory movements is still more regular. These movements always fall into oscillations of different orders superimposed on each other. In general, such a great complexity is to be seen in the manifestation of oscillatory movements during the course of their history that one must think not of the massive uplift and subsidence of large rigid blocks, but of a process resembling the action of a delicate and precise mechanism—which can be due only to the complex interplay of vertical forces, changing in direction and rate at various places, and not to simple warping of the crust by general forces of compression.

The most serious defects of the contraction hypothesis appeared, not surprisingly, in studies of the very problem it was designed to solve—that of the origin of folding. Domical folds on platforms are common among the folded dislocations. It will be seen that the formation of this type of folding cannot be explained by horizontal pressures. But even in folded zones, where linear folds have developed, phenomena are observed that exclude the action of exterior tangential forces. For example, the isolation of folded zones in more or less self-contained ovals (such as the Hungarian, Western Mediterranean, Caribbean, and many folded zones), surrounded by what the contractionists believe is a rigid platform, makes it impossible to understand the formation of folds within such an oval if the pressure that creates them is applied from without. In such cases there is no room for the discrete rigid parts of the platform to converge, since the parts which lie on both sides of the folded region form a single rigid body.

In the same manner, the action of external lateral forces is excluded in cases where the groups of folds terminate in blind ends surrounded like a horseshoe by the platform. Such cases are encountered in the northwest end of the Donbas, in both plunges of the Greater Caucasus, in the ends of the Pyrenees, in the terminations of a number of anticlinoria on the western slopes of the southern Urals, and in many other places. The formation of folds first in the middle part of the geosyncline and then spreading to the periphery agrees poorly with the contraction hypothesis, which is also contradicted by the transverse folds observed in the plunges of large anticlinoria.

It is quite clear that O. Yu. Shmidt's conception of the earth's development, the mechanism of compression of the crust, has no place, at least in relation to past geologic time, since Shmidt's theory assumes that the cooling phase has not yet set in.

From all these contradictions it follows that the contraction hypothesis cannot serve as a basis for tectonic generalizations, either as a general theory of the future gradual cooling of the earth or as a concrete explanation of the mechanism of tectonic movements. A tectonic theory cannot be built around the idea that horizontal forces of compression and horizontal displacement of the material of the crust are dominant

factors in the structural history of the earth's crust. Rejection of the contraction hypothesis was thus inevitable.

FORMATION OF THE RUSSIAN SCHOOL OF GEOTECTONICS

The great Russian scientist M. V. Lomonosov must be considered the founder of scientific geotectonics. The original theory developed by him was widely circulated among Russian scientists in the first half of the nineteenth century. An important work, D. Sokolov's *Course of Geognosy*, was published in 1839. This book examines the whole complex of geologic processes in the light of the properly understood interrelationships between external and internal processes—the fundamental meaning of “internal” being plutonic.

Russian investigators in the middle of the last century successfully developed the ideas of the uplift hypothesis. These ideas, as already mentioned, were held by E. Eykhval'd, G. D. Romanovskiy, N. A. Golovinskiy, G. Ye. Shchurovskiy, and others.

Later the contraction hypothesis came to be as widespread in Russia as in the West. Subsequently, however, in proportion as enthusiasm grew in Western Europe for the formal approach, the increasing isolation of the Russian geotectonic school could be noted. This isolation was reflected in the fact that the formalistic tendencies of the Western European tectonists received no support from leading Russian scientists, who were able to stay on the true course of the historical, evolutionary approach to the phenomena they were investigating.

The majority of Russian geologists studied the interrelationship and development of natural phenomena and never regarded description of the morphologic aspect of the phenomena as an end in itself. In the field of geotectonics, geologists studying the structure of the earth's crust did not forget its material composition and never put purely geometric analysis in the foreground. Earlier than anywhere else, they developed methods of investigating the material composition of the crust in order to reconstruct the history of its movements. This reconstruction, of course, included the oscillatory movements, which had at that time been almost forgotten by western students of geotectonics. The fact that the study of oscillatory movements received greater attention in Russia gave our geotectonics important advantages. The first clear concept of the course of oscillatory movements and its results in the nature of the formation of strata was suggested by Golovinskiy (1869).

The final molding of the Russian school of tectonics, with its historical tendencies, is associated with the name of A. P. Karpinskiy. Thus his works, which were devoted to the structure and history of the European part of Russia, are of particular interest (1883, 1887, 1894, 1919).

On the basis of his generalizations from many geologic explorations

of the Russian platform, Karpinskiy concluded that the granite-gneiss basement of the platform was cut up by faults into grabens and horsts. Such horsts are, for example, the Baltic and the Azov-Podolian shields. In the southern part of the platform he noted the existence of two lines of dislocation, with a west-northwest strike, which later were given the name *Karpinskiy's lines* in the literature. A. P. Karpinskiy first called attention to the dislocation of beds within the Russian platform, and he made the first attempt to explain their formation. On this question he had two views at different times. In one view, which paid some tribute to the compression hypothesis, he saw these dislocations as the result of pressures transmitted through the body of the Russian platform from the folded zone of the Urals. According to the other, which was entirely original, he considered that these dislocations were caused by the platform's own movements.

In another series of works (1894), Karpinskiy, after studying the facies of the deposits, produced paleogeographic maps of the Russian platform. A comparison of these led him to the following conclusion:

Within those parts of the earth's surface now occupied by European Russia there have been successive oscillations of the earth's crust, taking the form of an alternation between subsidence parallel to the equator and subsidence parallel to the meridian. These slow, or wave-like, oscillations encompassed all but the northwest part of Russia, where a block of the crust composed of the oldest crystalline formations and representing a so-called horst, acted as a bulwark or buffer, about which, as if it were a stationary axis, the movements of subsidence and uplift took place. . . . The oscillations have almost always proved to be parallel to the ranges of the Caucasus and the Urals. During the most intensive development of the Urals, subsidences parallel to the meridian occupied the greatest length of time; during the intensive development of the Caucasus, subsidence parallel to this range predominated in length of time.

A. P. Karpinskiy's paleogeographic conclusions represented, for his time, an enormous achievement, the significance of which extended far beyond the borders of Russia. Other attempts at similar generalizations then known in foreign literature applied only to particular cases. The investigators who made such attempts did not put to themselves or solve such broad problems as did Karpinskiy: to reconstruct the history of movements of the earth's crust over a great area and a very long period of time, on the basis of generalization from a huge amount of factual material.

It should be remembered that these paleogeographic studies of A. P. Karpinskiy's were published at a time of universal absorption in folded structures and ignorance of any tectonic movements other than folding, especially in regions outside geosynclines. A. P. Karpinskiy's work was in opposition to the prevailing formalistic tendencies. He made a be-

ginning in the methods of studying tectonic history from an analysis of the composition of rocks and laid the foundations of the doctrine of oscillatory movements of the earth's crust. He was the first to occupy himself with a thorough tectonic study of platforms, regarding them not as inert blocks but as developing geostructural elements.

Almost at the same time as Karpinskiy's first publications, there appeared a number of studies by A. P. Pavlov, devoted chiefly to the geologic structure of the Volga River area. These studies offered solutions to basic problems of the stratigraphy of the Mesozoic deposits in the central regions of the Russian platform and presented the first detailed analysis of the tectonic structure of these regions. Pavlov, for example, first explained the structure of the Zhiguli district and established the existence within the platform of very gently sloping tectonic depressions, for which the term *syncline* has been introduced (Pavlov, 1887). Thus A. P. Pavlov's work developed the ideas introduced by A. P. Karpinskiy of the platform as a mobile part of the earth's crust.

Among the achievements of Russian geotectonics before the 1917 Revolution, mention should be made of the schematization of the structure of central Siberia, as published by I. D. Cherskiy (1886). This was used by Suess, who incorporated it into his general tectonic survey of the earth.

At the end of the nineteenth century a detailed survey of the Donets coal basin was made by a group of geologists under the direction of L. I. Lutugin. The maps drawn by this group reveal an excellent technique in cartography combined with a thorough understanding of the geometry of the structure. By these qualities the detailed maps of the Donets basin are in no way inferior to the best examples of Western European maps of that time.

THE PRESENT STATE OF GEOTECTONICS

Some geologists found themselves in a state of confusion when the contraction theory, which they believed to be entirely true, began to reveal its inadequacy. Under the influence of this disappointment, some researchers lost faith in the possibility of finding answers to the fundamental questions in the structural development of the earth's crust, and they began to retreat from such problems. They devoted themselves to studying particular tectonic processes, considered in isolation, and to investigating the detailed morphologic features of various structural forms.

Other geologists tried to overcome the crisis in ideas by creating new theories and hypotheses, which in many respects contradicted each other. The authors of the new hypotheses introduced to replace the contraction hypothesis frequently tried to correct those deficiencies in

the latter which were discovered in solving the problems of oscillatory movements—above all, that of their reversibility. This was the course taken by Bucher, who proposed the pulsation hypothesis (1933). According to this hypothesis, stages of universal expansion and contraction alternate in the earth's history. Expansion produces subsidence of the weaker parts of the earth's crust and the formation of geosynclines, whereas contraction causes folding and uplift.

This was, in its time, an interesting new attempt to explain the complexity of the processes occurring in the earth's crust. The pulsation hypothesis had most of the shortcomings of the contraction hypothesis and added a number of its own. In the history of the earth we find no epoch of general expansion. Not only does the subsidence of geosynclines at the beginning of the cycle not occur at the same time in different places; even within a single geosyncline subsidence is always closely interwoven with uplift. In the second stage of its development, the geosyncline is usually divided into a central area of uplift and peripheral zones of subsidence. In the central region the uplift takes place simultaneously with folding. According to the pulsation hypothesis, folding and uplift are produced by contraction and subsidence by expansion. But this leads to an impossibility, since we are forced to assume that the area of contraction is located within the area of expansion.

Investigators have tried to eliminate the same deficiency in the contraction hypothesis, which appeared in connection with the problem of oscillatory movements, by completely different means. The suggestion has been made that folding and oscillatory movements are two categories entirely independent of each other; although folding is caused by compression, vertical movements may be of a different nature and may be the reaction of the earth's crust to increase or decrease in the static load of sedimentary rocks (Dutton, 1892).

Apart from the fact that the theory of isostasy has not been able to explain the complex process of oscillatory movements, the formulation of the problem itself is hopeless. Our goal, it has been pointed out, should be to establish the causal relationships among phenomena. Here, on the other hand, an attempt has been made to divorce two fundamental phenomena artificially from each other. This is all the more unjustified in that at each step we are persuaded of the undoubted existence of very strong connections between these phenomena.

At the height of the crisis, when it seemed that all simple measures had been attempted unsuccessfully, many hypotheses appeared that reflected a further deepening of the formalistic tendency in geotectonics. These extremely diverse hypotheses were alike in one respect, that they were all equally removed from reality. They were, as a rule, based on arbitrary assumptions, and by schematizing geologic history, they distorted it. In some cases these hypotheses were formulated with great

literary skill, so that they played on the imagination and achieved wide circulation. They have caused considerable damage to geotectonics, giving nonspecialists the impression that this is a field in which the most superficially conceived fantasy reigns. The clearest example is Wegener's hypothesis of continental drift (1924). This both contradicts the geologic facts and fails to explain the things that most require explanation.

A close look, however, reveals that even then hypotheses of this sort did not constitute the whole of the science. Work based on actual geologic data continued; as a result, a new trend appeared in geotectonics that provided a way out of the crisis and indicated the direction in which the science might develop further. The basis of this trend was the idea of the predominance of vertical forces in the earth's crust. This is, in essence, a return to the fundamental principle on which geotectonics was based one hundred years ago; this principle has, of course, been given a completely new formulation. What was the reason for this strange-seeming vacillation in geologic thinking and its apparent return to the former concept of coordinated movements?

The neglect with which the contraction hypothesis had treated oscillatory movements turned out to be unjustified. It was discovered that oscillatory movements of the earth's crust are a complex and abundant, as well as ubiquitous and constant, phenomenon, forming the background of other types of local and intermittent tectonic movements. When this became clear, a return to the old idea of the fundamental importance of vertical forces and movements was inevitable.

The first modern attempt to derive general geotectonic conceptions from the idea that vertical oscillatory movements of the earth's crust are primary was made by Ampferer (1906), who suggested that the vertical movements of the crust are caused by vertical currents in the igneous rock mass below the crust. In flowing toward the sides, this mass draws the crust after itself and crumples it into folds.

A second attempt, based on ideas proposed long before by Reyer (1888), was undertaken by Haarmann in his "oscillation theory," which saw folding as the result of the slippage of stratified rocks under the force of gravity down the slopes of uplifts created by vertical forces (1930). This attempt, which was refined by van Bemmelen (1933, 1955), established the "gravitational hypothesis" in present-day geotectonics.

The third attempt, finally, was that of M. M. Tetyayev, who suggested that folding might be the result of the vertical compression of beds during the vertical movements of the earth's crust (1934, 1941).

The main features of the prerevolutionary Russian school of geotectonics have been described above. After the October Revolution there was radical change in the development of geotectonics in Russia. The

growth of the socialist planned economy made increasing demands on geology. During that time geology acquired new capabilities, which had been unforeseen in prerevolutionary Russia, resulting from the enormous increase in the extent and scope of the various geologic investigations and the organization of scientific research institutes, etc.

The new situation created the soil for the rapid growth of Russian geotectonics as an independent science. On the one hand, it had the opportunity to use new, exceptionally rich, and varied material—the results of geologic observations over the extensive territory of the Soviet Union. On the other hand, the conditions were created for generalizing from this material through the organization of special scientific research work. The increased feeling of responsibility on the part of geologists for solving diverse structural and tectonic problems for practical purposes contributed greatly to the recent achievements of geotectonics in the Soviet Union. In the great variety of the practical problems confronting geology, there has come to be a practical urgency for solving many problems that not long ago seemed academic.

The following characteristics of Soviet geotectonics have become especially clear in recent years:

1. A scrupulous adherence to factual material and hostility to the creation of speculative syntheses.

2. A historical approach to the study of phenomena as the basis for scientific analysis.

3. A conception of unity in the development of the earth's structure and of the interrelatedness and interdependence of its various aspects.

4. A concept of the universality of the tectonic process, in which tectogenesis is considered as the development of the entire earth and, in particular, of the entire crust, each portion of which participates in this development both independently and as a part of the total process. In these concepts there is no room for inert rigid blocks or plastic zones crumpled by external forces that are extraneous to them. There is a conviction that tectonic forces are present everywhere, under every part of the earth's crust, and that their primary direction is vertical. From all this comes the idea that oscillatory movements are the main form of the tectonic process.

5. A striving to derive theoretical generalizations from data obtained over extensive territories.

6. A search for new methods of investigations.

Each of the points enumerated deserves brief comment. The first years of the growth of Soviet geotectonics coincided with the time in which the literature of other countries was flooded with a huge number of speculative hypotheses. This was the period immediately after the collapse of the hypothesis of contraction of the earth's crust. Soviet researchers followed the foreign experiment attentively. At the beginning,

interest took the form of critical reviews of foreign literature. These included reviews by A. A. Borisyak (1924), Ye. V. Milanovskiy (1923, 1924, 1929), V. A. Obruchev (1926), and N. M. Strakhov (1932), who not only reported the opinions of the foreign authors but also posed new questions. Some Soviet researchers, moreover, were not immune to the attraction of the foreign ideas. The hypotheses of Wegener and Joly and the ideas of Kober, Stille, and Bubnoff were at one time circulated and accepted by some Soviet geologists. This enthusiasm died quickly, and since the thirties of this century such theories have had no success in Russia. The most typical feature of Soviet theoretical geotectonics is the extensive use of empirical laws, especially those which relate the various types of tectonic movement to each other; for example, oscillatory movements to folding, and these dislocations and others to fracturing.

The historical approach to the study of tectonic phenomena, which is clearly reflected in A. P. Karpinskiy's works, was still more strongly emphasized in the Soviet era. Soviet geologists have almost never been interested in the purely formal description of structures. In studying folded structures, for example, Soviet investigators turned their chief attention to discovering the laws in the development of these structures.

Soviet geologists have elaborated the concept of unity in the development of the earth's structure. They have been firmly convinced that all manifestations of the earth's tectonic life are closely and regularly interrelated, together forming an integral process that merely changes its form in time and space. Soviet geologists have therefore avoided the "dualistic" theories of foreign authors, who are, for instance, trying to explain folding and oscillatory movements by entirely different and independent causes. M. M. Tetyayev deserves much of the credit for developing the concept of the unity of tectonic processes.

An understanding of the unity of geologic phenomena makes it possible to study the relationship between tectonic processes and the surface processes. This outlook has already proved to have a favorable influence, for example, on the scientific treatment of sedimentation and sedimentary formations.

The interest in oscillatory movements is closely associated with the attempt of Soviet scientists to approach tectonic phenomena from the point of view of their development. Oscillatory movements in particular are "historical"; since they are universal and continuous, their history most clearly reveals the laws governing the development of the earth's structure.

Methods of studying the history of ancient oscillatory movements, based on analysis of the thicknesses and the facies of sedimentary rock series, have been worked out with the greatest thoroughness in the U.S.S.R. These methods were first applied to a study of the develop-

ment of large territories over long periods of time. The result has been a new theory of oscillatory movements.

The complex picture of oscillatory movements revealed by these new methods is not in harmony with the hypothesis of contraction of the earth's crust. It leads, rather, to the conviction that the fundamental and primary direction of the tectonic forces in the earth's crust is vertical and that the visible horizontal displacements should be viewed as secondary, derivative phenomena. The supremacy of vertical movements may be considered as established. The next problem is to determine how vertical movements cause the formation of folds. We have seen that it was precisely at this point that the old hypothesis of uplift met with failure in its time.

Domical folding on platforms is directly connected with vertical forces. A dome is the result of local vertical forces of uplift. The difficulties arise in explaining the origin of linear folding, of the elongated wavelike folds that lie en echelon within the folded zone. The connection between vertical forces and crumpling of strata into folds and the means by which vertical stresses are transformed into horizontal movement of the material in the earth's crust is one of the most thorny problems of present-day geotectonics. The magnitude of the vertical and horizontal movements in the earth's crust does not, in principle, contradict the possibility that there is such a transformation and that the horizontal movements are subordinate to the vertical. The old hypothesis of uplift was too simple to serve as a means of answering this question. Now, of course, this problem requires a deeper and more thorough solution; how such a solution may be found will be discussed below.

It must be stressed that, in maintaining the view of the preeminence of vertical forces, we are radically changing the concepts of the mechanics of the earth's crust that had accompanied the contraction hypothesis. Its adherents had assumed a uniform distribution of the forces of horizontal compression, so in order to explain the differences in the tectonic process as it occurs in various parts of the earth, they inevitably had to regard the mechanical properties of the crust as essentially heterogeneous.

In the new view, the mechanical state of the earth's crust, or of its separate parts, loses its decisive importance: the differences in the course of the tectonic process depend to a greater degree on the non-uniformity and the unequal intensity of the vertical forces acting directly upon each part of the crust.

Furthermore, the problem of the location of the centers of the tectonic activity that causes folding and all the other tectonic phenomena in geosynclines is solved on a different principle. According to the contraction hypothesis, this activity is concentrated outside the limits of the

geosynclinal zones and takes the form of pressure from the "rigid" platforms. It was not by chance that Kober called the platforms *kratogens*—sources of power. In the new conception the process that produces folding, as well as igneous activity and the more intensive oscillatory movements in geosynclines, arises within the geosynclines themselves, which are not areas reacting passively to pressure from outside but the chief centers of tectonic activity. In particular, the horizontal movement of the masses, which leads to the formation of folds and is directed from the center of the geosyncline toward the periphery, arises not because the adjacent platforms crush the geosynclines in moving toward each other but because movement of material in this direction begins, when the platforms are relatively immobile, within the geosyncline itself. With this view one can surmount the difficulties encountered by the adherents of the contraction hypothesis in explaining the mechanisms of folding, which were raised by the shape of the folded zones and the presence of transverse folds, etc.

If folding results from the action of local forces, a special significance attaches to an investigation of the historical prerequisites for the origin of folding—of the oscillatory movements that precede it and the resulting distribution of rocks with different compositions. The new theory has undoubtedly freed geotectonics from its formalistic tendencies.

Materials for theoretical generalizations in the field of geotectonics are to be found in the regional tectonic surveys compiled by Soviet geologists in recent years for extensive territories of the U.S.S.R. Examples of excellent studies of regional tectonics are the well-known works on the structure of the Russian platform by A. D. Arkhangel'skiy (1923, 1932, 1934, 1941, 1947–1948), who continued A. P. Karpinskiy's classic investigations. In his latest surveys A. D. Arkhangel'skiy considered the structure of the entire territory of the U.S.S.R. The studies of the regional tectonics of Siberia begun by I. D. Cherskiy were continued after the October Revolution by V. A. Obruchev (1927, 1935–1938). This trend toward regional studies is reflected in N. S. Shatskiy's publications (1932, 1937, 1945, etc.). The tectonics of geosynclinal regions were studied particularly by D. V. Nalivkin, V. P. Rengarten, V. I. Popov, V. Ye. Khain, V. V. Belousov, N. A. Shtreys, M. V. Muratov, A. V. Peyve, A. A. Bogdanov, and other researchers.

Since 1945 many works have appeared on the analysis of the structure and tectonic history of the major individual regions of the U.S.S.R. The construction of a Tectonic Map of the U.S.S.R. has been an important achievement.

The search for new methods of investigation is seen in Soviet publications; for example, in the study of oscillatory movements. These methods will be discussed in the appropriate section of this book. Innovations in method also appear in the extensive application of geophysical methods

to the study of tectonic structure. An interpretation of the structure of the basement of the Russian platform, which lies at great depth, was made by A. D. Arkhangel'skiy, N. S. Shatskiy, and V. V. Fedynskiy on the basis of geophysical data; in many respects this is in principle an achievement of great importance.

By gathering a great amount of factual material and elaborating new methods of analyzing this material, Soviet geologists have been able to make a beginning in deriving broad generalizations in the field of geotectonics, in solving its fundamental problems, and in postulating a theory to explain the trend and the basic laws in the development of the earth's crust. The present state of these problems and the basic ideas in this field will be discussed at the end of this book.

On their way to constructing a general theory, Soviet geologists are working intensively on many intermediate problems. These involve a study of the various tectonic phenomena, particularly the establishment of numerous partial causal connections and relationships between them. As an example, the association between oscillatory and folding movements is being investigated to establish regularities showing how the regime of folding changes with changes in the regime of oscillatory movements.

The laws governing the periodicity of tectonic movements, the association between oscillatory movements and folding, the relationship of these and other movements to the formation of fissures and faults and to igneous activity have all been studied to such a degree that basing general predictions of structure on these regularities does not present insurmountable difficulties.

To make such predictions of the general structure specific and more detailed, so that they may be applied to practical work, studies of the mechanism in the formation of intermediate and small-scale structures have been begun in the U.S.S.R. in recent years. Important results have been achieved in determining the conditions for the formation of cleavage, tectonic fissures, orientation of the texture, small disharmonic folds, etc. The systematic use of the methods of physical experiment on scale models in studying the mechanism of the formation of various structures was begun in the U.S.S.R.; this has been the basis for the development of a new branch of science—tectonophysics.

PART TWO

THE INTERNAL STRUCTURE
AND COMPOSITION OF THE EARTH.
THE ORIGIN OF THE EARTH

CHAPTER 3

STRUCTURE AND COMPOSITION OF THE EARTH'S INTERIOR. THE ENERGY AND AGE OF THE EARTH

THE SHAPE AND SIZE OF THE EARTH

The earth is a spherical body composed of solid, liquid, and gaseous substances distributed in a definite sequence. The central, main part of the planet is made up of solid matter; its outer shell is called the *lithosphere*. The surface of the lithosphere is covered by an envelope of water, the *hydrosphere*, which is not continuous but contains extensive gaps. The outermost portion of the planet is a gaseous envelope, the *atmosphere*.

In determining the shape and dimensions of the earth, we regard the surface as the surface of the hydrosphere, projected through the continents and everywhere normal to the plumb line. The true shape of the earth is extremely complicated and does not correspond to any form subject to mathematical analysis. This complex shape has been named the *geoid*. The concept of the earth's shape can, however, be simplified. The geoid differs only slightly from an *ellipsoid of revolution*, or *biaxial ellipsoid*, which is somewhat flattened at the poles. For this reason, the equatorial and polar radii of the ellipsoid are not equal. According to computations made by F. N. Krasovsky, the polar radius of the earth c is 6,356,863 m, whereas the average equatorial radius a is 6,378,245 m. The equatorial section is also somewhat flattened, like the meridional, but to a very slight extent. The equatorial radius through the meridian 15° east of Greenwich is approximately 200 m longer than the one perpendicular to it. Thus the earth most closely approaches a *triaxial ellipsoid*; this, however, will be disregarded in the following discussion. The ratio of the difference between the average equatorial and the polar radii to the longer of the two, or $(a - c)/c$, is the oblateness of the earth; this is approximately $\frac{1}{298}$. If one were to build a globe with an average radius of 1 m, the difference between the two radii would be only 3 mm. This small degree of oblateness allows us to consider the earth as a sphere when no particular precision is required. In such a

case, the radius of the earth can be taken as approximately 6,370 km. Its surface area is $5.1 \times 10^8 \text{ km}^2$; its volume is $1.08 \times 10^{12} \text{ km}^3$; its mass is $5.98 \times 10^{27} \text{ g}$ [$5.98 \times 10^{27} \text{ g}^2\text{-Ed.}$]

The depth of the hydrosphere varies from zero to 10,990 m (in the deepest ocean trough, near the Mariana Islands). The average depth of the hydrosphere is 3,800 m.

According to the latest data, the height of the atmosphere reaches 1,000 km; this is the greatest altitude at which physical phenomena indicating the presence of gas have been observed.

The mean altitude of the dry land above sea level is 850 m; the highest point of the lithosphere is about 8,900 m above sea level. The greatest variation in the relief of the earth's surface is therefore no more than 20 km, or $\frac{1}{300}$ of the radius of the earth.

A brief survey will be presented below of the existing information on the structure of the earth's interior, as obtained by various geophysical methods. For a detailed discussion, see the primary sources: Poldervaart, 1955; Ahrens, 1956; Gutenberg, 1958; Kropotkin et al., 1958; Magnitskiy, 1953; Kosminskaya, 1958; Demenitskaya, 1958; Lyubimova, 1958; Gutenberg, 1951; Heiskanen and Vening-Meinesz, 1958; Gutenberg, 1957; Ewing and Press, 1955.

THE DISTRIBUTION OF DENSITIES WITHIN THE EARTH

It is well known that the mean density of the earth, as computed from its total mass and volume, is 5.52. The rocks near the surface, however, have much lower densities. The mean density of sedimentary rocks may be taken as 2.3, and that of the entire earth's crust as a whole is considered to be 2.7. The interior of the planet must, therefore, have a density much greater than the mean density of the entire earth.

Table 1. The Distribution of Densities with Depth (After Bullen)

Depth, km.....	33	410	1,000	2,900 and above	2,900 and below	4,980	6,370
Density.....	3.32	4.07	4.41	5.57	9.7	12.10	17.9

The most likely distribution of densities in depth has been determined by a combination of astronomic and seismic methods; these include calculations of the earth's shape, the peculiarities of its motion as a planet, the migration of the poles, and the propagation velocity of seismic waves at different depths. This distribution of densities is shown in Table 1.

Gravimetric measurements at the earth's surface have shown that the surface relief does not have the effect on the force of gravity that it

should have if the density of the earth's crust did not change in the horizontal direction and if the bottom of the crust were horizontal. If the latter condition were actually the case, the mountains would represent additional masses and the force of gravity near them would be greater than in the plains. The ocean basins, on the other hand, would be areas with a smaller force of gravity, since here the greater mass of the rocks would be replaced by the smaller total mass of the water. In actual fact, this is not the case: it may be said, to the first approximation, that the force of gravity is the same over the whole of the earth's surface. This will be true in every case if the forces of gravity to be compared are averaged over great areas of the surface—several hundred kilometers in radius. Thus the conclusion is inevitable that areas in which the relief of the earth's surface is convex are compensated by lighter masses at depth, whereas beneath the concavities in the surface (such as the ocean basins) there must be heavier masses at the same depths. The result is that the total mass summed along a vertical line is almost the same everywhere.

The undoubted fact that mountains are compensated by lighter masses at depth has produced the idea of the so-called mountain roots, the existence of which, as will be seen below, has recently been confirmed by direct seismic evidence. "Mountain roots" are protrusions of the upper layer of lighter material downward into the layer of heavier material, which is driven to the side, thus producing the necessary compensation. An extremely simplified calculation may be given as an example: the density of the earth's crust is 2.7, and the density of the upper part of the mantle is 3.3. The earth's surface protrudes 4 km into the atmosphere at a certain point; to compensate for this excess mass at the surface, the material of the crust must project into the material of the mantle below this protrusion. Since the difference between the densities of the crust and mantle is 0.6 and since 2.7 divided by 0.6 is equal to 4.5, to compensate for the protrusion of the earth's surface the root of the crust must project $4 \times 4.5 = 18$ km into the mantle. Similar calculations will show that to compensate for a depression in the earth's surface 2 km in depth, for example, the surface of the mantle must rise 9 km above its normal depth.

Since, to the first approximation, the mass of the earth's crust beneath any point on its surface is the same as beneath any other point, the crust is, in general, in a state of equilibrium. Such an equilibrium would have to be established, according to Archimedes' law, if the earth's crust were floating on a denser but still viscous substratum. Let us imagine a nonhomogeneous crust floating on the surface of a substratum that is denser but fluid enough that the crust may sink into it, in conformity with Archimedes' law. This law states that a floating body displaces an amount of the substance in which it is floating

equal to its own weight. Therefore more of the substratum is displaced under the heavier parts of the crust than under the lighter, and the former sink to a greater depth than the latter. Since every part of the earth's crust displaces only an amount of substance equal to its own weight, the presence of a part of the crust, of any given density and thickness, floating at a point on the earth's surface does not change the total mass of material beneath this point. This equilibrium of the earth's crust is known as *isostasy*.

The existence of isostasy indicates that the substance in the earth's interior behaves as a viscous fluid, which is capable of being displaced, depending on the inequality of the loads it carries on its surface. Such displacements are, of course, very slow, so a long time is required for equilibrium to be established. Within the limits of geologic time, however, equilibrium may be attained. This ability of a solid body to undergo slow plastic deformation when a load is applied to it over a long period of time is called *creep* (see Chap. 18 below). Thus the earth's interior has the property of creep.

The fact that the terrestrial sphere as a whole behaves as a plastic body, viewed in the scale of geologic and astronomic time, may be seen from the shape of the earth. It has been calculated that the degree of oblateness of this planet is exactly what it should be if the planet were composed of fluid matter.

The earth's isostatic equilibrium, however, is observable only when one considers the average magnitudes of the force of gravity for large portions of the earth's surface, with a radius of several hundred kilometers. A more detailed study of the gravimetric field in particular areas of the earth will reveal considerable deviations from the state of equilibrium, in the form of positive or negative gravitational anomalies. The greatest anomalies have been observed in zones of the most intensive tectonic movement. Moreover, it has often been determined that the present-day movement of the earth's crust (uplift or subsidence) in such zones is in the direction opposite to that of the forces tending to restore isostatic equilibrium. The main Caucasus range, for example, has a certain mass in excess of the mass required to maintain its isostatic equilibrium. To attain this equilibrium, the Caucasus would have to subside; instead of this, it is rising.

Many similar cases provide convincing evidence that isostatic forces—the tendency to restore isostatic equilibrium—cannot be considered the cause of tectonic movements of the earth's crust. Tectonic movements, on the contrary, destroy the equilibrium, as in the example of the Caucasus. The constant attempt of the earth's crust to maintain equilibrium may be regarded as a regulating factor, or brake, on tectonic movements. When the tectonic movements that upset the equilibrium are rapid, the establishment of a new isostatic equilibrium is always

retarded; if tectonic movement takes place very slowly, the restoration of isostatic equilibrium by appropriate displacements of the material in the earth's depths may occur almost simultaneously. Ye. N. Lyustikh and V. A. Magnitskiy have shown, for example, that the slow vertical movements on platforms that result in the formation of antecises and synecises (to be discussed below) are accompanied by horizontal displacement of the material in the depths—moving outward from beneath a synecise and inward beneath an antecise.

Gravitational anomalies are extensively used in applied geophysics, in studying the structure of the parts of the earth's crust close to its surface.

THE SOLIDITY OF THE EARTH

Although it has just been shown that the material of the earth, when subjected to a load of long duration, behaves as a plastic body, and although it possesses great viscosity, mechanical action of short duration upon the same material produces elastic vibrations whose energy and velocity of propagation indicate that the earth is "solid"; that is, that its material has a modulus of rigidity. This conclusion also follows from a study of the "solid flow" in the earth's crust—of the waves produced in the surface of the crust by the gravitational attraction of the moon and the sun. Table 2 shows the modulus of rigidity as determined by seismic methods for several depths.

Table 2. Modulus of Rigidity of the Earth's Substance (After Bullen)

Depth, km.....	640	1,900	2,900
Modulus of rigidity, dyne/cm ²	1.4×10^{12}	2.4×10^{12}	3.0×10^{12}

The modulus of rigidity of certain materials may be given for comparison:

Granite	2.0×10^{11}	Iron	5.1×10^{11}
Gabbro	4.4×10^{11}	Steel	7.9×10^{11}

The fact that a single material may have plastic properties and at the same time be more rigid than steel should cause no surprise. The mechanical properties of a body depend to a great degree on the velocity of the mechanical action to which it is subjected: ice, for instance, will vibrate after a strong, rapid blow, but under the long-lasting action of the force of gravity it will flow downward along mountain valleys as glaciers. In exactly the same manner, the substance of the earth is plastic in relation to mechanical action of long duration and extremely rigid—that is, capable of very great resistance to changes in its shape—under the influence of brief mechanical action such as elastic vibrations.

PRESSURE WITHIN THE EARTH

Because of the weight of the overlying masses, the pressure increases as one moves toward the earth's center. Table 3 shows a few pressures at certain depths, as computed by Bullen.

Table 3. Pressure within the Earth

Depth, km.....	200	1,000	2,900	4,980	6,370
Pressure, 10^{12} dyne/cm ²	0.07	0.40	1.33	3.22	3.94

THE EARTH'S THERMAL REGIME

The daily and annual variations in the temperature of the earth's surface penetrate only a small distance into the earth. At a depth of usually no more than 20 to 25 m, there is a level of constant temperature equal to the mean annual temperature of the given locality. As one penetrates deeper into the earth, the temperature rises (except in regions of permanent frost, where low temperatures go down to great depths). This increase in the earth's temperature with increasing depth, measured per unit length, is called the *geothermal gradient*; its reciprocal, the distance that must be penetrated to observe an increase in temperature of one degree centigrade, is called the *geothermic degree*. The magnitudes of the geothermal gradient and the geothermic degree will differ in different places. The smallest geothermic degree, 6.7 m, has been recorded in the state of Oregon, U.S.A., and the greatest, 434.3 m, in South Africa. The average geothermal gradient for the entire earth is 3×10^{-4} deg/cm, which corresponds to a geothermic degree of 30 m.

From the geothermal gradient and the thermal conductivity of the rocks one may easily compute the amount of heat lost by the earth. The average heat flux at the earth's surface may be taken to be 1×10^{-6} cal/cm²/sec; individual values may vary between the limits of 0.58 and 3.3×10^{-6} cal/cm²/sec. The total heat transfer rate from the earth is equal to 9.2×10^{12} cal/sec.

At the present time there is no doubt that the heat emitted from the earth is primarily of radioactive origin; this heat is given off in the decay of the radioactive elements contained in all earth materials. Data on the heat given off by some typical rocks are presented in Table 4, which shows that basic rocks emit considerably less heat than acidic rocks.

In addition, it is of interest to determine the amount of heat transfer on the continents and on the ocean bottom; a comparison shows that this is approximately the same in both bases. On the continents, however, the earth's crust contains a thick layer of granites, whereas be-

neath the oceans the granites are absent and the crust is composed only of basalts. The equality in the heat transferred in both cases can thus mean only that the average composition of the material in the earth's crust, taken over a very great depth, is similar in each case but that the distribution of this material differs. Beneath the continents, more radioactive material is concentrated in the granites close to the surface; beneath the oceans, this material is dispersed through a much greater thickness of the earth's crust.

Table 4. Amounts of Radioactive Heat Given Off by Typical Rocks

Rock	Contents of		Emission of heat, cal/g $\times 10^6$ yr			
	U, 10^{-4} %	K, %	U	U-Th	K	Total
Granite	4	3.5	3	6	1	7
Intermediate rocks	2	2	1.5	3	0.5	3.5
Basalts and gabbros	0.6	1	0.4	0.8	0.3	1.1

The geothermal gradient must decrease with depth: if the temperature increased down to the center of the earth at the same rate at which it increases near the surface, the interior of the earth would have such a high temperature that the planet could not exist as a solid body. It has been conjectured that the earth's temperature must be as high as 150 to 180° at the depth of 10 km, increasing to 230 to 370° at 20 km and to 300 to 500° at 300 km. Below 20 km, the temperature must increase very slowly to several thousand degrees at the center of the earth. At the present time the most likely temperature at the center of the earth is thought to be 4000°.

Ye. A. Lyubimova (1958) has attempted a quantitative reconstruction of the history of the earth's temperature, basing her calculations on O. Yo. Schmidt's theory (1950) of the origin of the planet. Her study indicates that, because of the very low thermal conductivity of the earth, almost no heat flows outward from depths below 1,000 km. The continuing radioactive decay therefore causes a slow gradual increase in the temperature of the center of the planet (about 200° per one billion years). At the same time, heat is being emitted from the surface of the earth at the rate of 100° every billion years. The maximum heat transfer from the surface took place approximately 1.6 billion years ago. The present author suggests that Lyubimova's conclusion may agree closely with what is known of the behavior of the earth's crust in Archean times, when the crust was characterized by a high degree of plasticity and processes of metamorphism took place very close to its surface. This may have been due to the high temperature of the earth's crust.

Because of the decay of the radioactive elements and the decrease

Origin, Internal Composition, Structure of the Earth

their total amount, the earth must ultimately cool off. Nevertheless, the quantity of radioactive elements in the earth is still very great, the thermal conductivity is very small, and cooling may therefore begin only after about ten billion years. At the present time only the surface layer is cooling, whereas the interior of the earth is still being heated.

The curve of the most probable distribution of temperatures in the earth shows that the substance between the depths of 100 and 700 km may be in a molten state. This agrees with certain conclusions drawn from seismic investigations of the earth's interior (see below).

If all these calculations are correct, the earth must be expanding at the present time. This expansion, however, amounts to an increase in the earth's radius of only 1 cm per 1,000 years and in its circumference no more than 10 m per million years. If it is assumed that the Paleozoic era began 500 million years ago, the earth's circumference must have increased by only 30 km during all this time. This is a negligible expansion, which could easily have been compensated by a very small amount of fissuring of the crust.

One interesting aspect revealed by these calculations of the earth's temperature is the opposition between the outer layers of the earth and its interior. The interior is expanding, whereas the outer layers, in cooling, oppose this expansion and therefore constrict the interior.¹

THE STRUCTURE OF THE EARTH ACCORDING TO SEISMIC DATA

Seismic methods are the principal means of investigating the structure of the deeper parts of the earth, which are not subject to direct observation. These amount to studying the paths and propagation velocities of elastic vibrations within the earth—the seismic waves produced by earthquakes. When the vibrations caused by distant earthquakes are registered on a seismograph, three kinds of waves are recorded: (1) *longitudinal waves* (similar to sound waves), propagating in solid, liquid, and gaseous media; (2) *transverse waves*, propagating only in solid media and related to the capacity of bodies through which they move to resist any change in their shape; and (3) *surface waves*, at the boundaries between solid, liquid, and gaseous media (their intensity rapidly decreases with depth). The longitudinal and transverse waves are sometimes called *forerunners*, since they arrive before the surface waves. During the passage of a longitudinal wave, the rock particles vibrate back and forth in the direction of wave propagation, generating in the propagating medium rays of alternating compression and rarefaction that move in all directions from the earthquake's focus. During the passage of the transverse waves, the particles vibrate in planes

¹ Rocks have very small strength in tension; hence the constrictive effect is slight.

perpendicular to the direction of propagation. Longitudinal waves are designated by the letter *P* and transverse waves by *S*.

The propagation velocity of longitudinal seismic waves is given by the formula

$$V_P = \sqrt{\frac{k + \frac{4}{3}\mu}{\rho}}$$

where *k* = bulk modulus

μ = modulus of rigidity

ρ = density

The propagation velocity of the transverse waves is computed by the formula

$$V_S = \sqrt{\frac{\mu}{\rho}}$$

If the earth's globe were homogeneous, consisting of materials with the same properties throughout, the propagation velocity of the waves would be the same everywhere and their path would be straight. In that case, the relationship between the time *t* during which the wave travels and the distance along the earth's surface *Θ* could be expressed by the formula

$$t = \frac{2R}{V_0} \sin \frac{\Theta}{2}$$

where *t* = wave travel time

R = radius of the earth

*V*₀ = wave propagation velocity in a homogeneous earth

Θ = angular distance, degrees of arc

Observations have shown that the true relationship is different: the waves reach distant stations sooner than they would if the earth were homogeneous; and the greater the distance, the greater the acceleration. It follows from this that the waves arriving at the more distant stations have been traveling faster. Since seismic waves travel not only along the surface but also through the depths of the earth, it is obvious that, because of the earth's curvature, stations more distant from the earthquake's focus will receive the waves that have passed through greater depths in the earth. Accordingly, the velocity of seismic wave propagation increases with depth. This suggests some change in the physical properties of the earth's substance. For instance, it has been established that at the depth of about 50 km the longitudinal waves travel at a velocity of 8 km/sec, whereas at the depth of about 2,900 km the velocity increases to 13.5 km/sec.

The increase in the propagation velocity of elastic vibrations with

depth is the reason for the curvature in the paths (rays) of the longitudinal and transverse waves; the paths appear to be convex toward the center of the earth. Therefore, after having reached a certain depth, a wave returns to the surface, where it is partially reflected and travels back into the earth, only to return to the surface once more and be reflected again, and so forth. These single and multiple reflections are designated by the symbols *PP*, *PPP*, etc., and *SS*, *SSS*, etc. For each of these waves, a so-called hodograph can be drawn. This is a time-distance curve for the relationship between the time in which a wave travels between the seismic focus and a given point and the distance it travels between them.

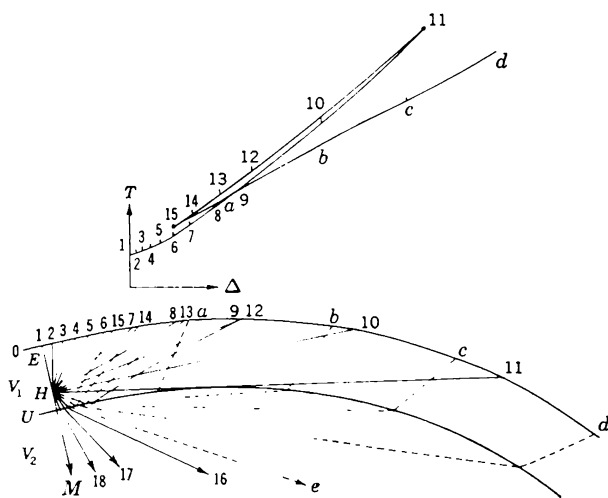


FIG. 6. Paths of seismic waves and their hodograph for an abrupt increase in wave velocity in a given layer within the earth.

Particular complications arise when the wave propagation velocity changes abruptly, at some depth. Two instances may be considered here. Let us assume that the velocity of a wave in a certain layer within the earth increases abruptly. The picture of the wave propagation under such conditions is given by the diagram in Fig. 6. It is assumed, for the sake of simplicity, that the wave velocity in the upper layer $O-U$ is constant; in the lower layer, below surface U , the velocity is also constant, but greater than in the upper layer. Wave paths 1-18 radiate from the focus H . Paths 1-11 are straight; paths 12-16 are reflected by boundary surface U ; paths $a-d$ are refracted. The last four waves travel along more complicated paths; but they travel faster than the straight rays 10-11, because they pass through a medium with a greater velocity of wave propagation. It may be seen that paths 15, a , b , c , and d differ only in that their segments lie below surface U . The

segments within the upper layer are almost equal. In the case of the refracted waves, the hodograph segments 15-a-d should be almost a straight line, because the parts of the waves below the surface of separation do not penetrate deeply. The corresponding hodograph is shown in the upper part of Fig. 6.

Let us now assume that the wave velocity decreases abruptly at a certain depth within the earth and that in the upper layer it increases with depth, while in the lower layer it is constant (Fig. 7). Then path 2 is tangent to the surface of separation. The next path is refracted and comes to the surface at point 3. The following waves emerge between points 3 and 4, whereas those with a still greater angle of incidence emerge at points 5 and 6. The path perpendicular to the boundary between the layers will pass through the earth in a straight line. It should be noted that not a single wave hits the surface between points 2 and 4; between them a peculiar seismic "shadow zone" is observed.

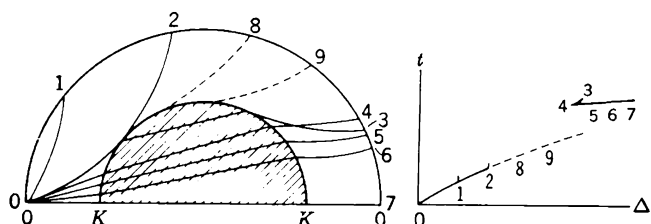


FIG. 7. Paths of seismic waves and their hodograph for an abrupt decrease in wave velocity in a given layer within the earth.

The corresponding hodograph is broken into two segments, as shown to the right in Fig. 7.

The above examples show that from the hodograph's appearance one may determine how the wave velocity within the earth changes with the change in depth. To do this, one must use data on the same earthquake, recorded by seismic stations scattered all over the earth's surface. From a comparison of such data, the distribution of the velocities of longitudinal and transverse waves with depth has been established; this is given in Table 5.

In this distribution of the velocities of seismic waves with depth, there are three surfaces, or discontinuities, at which the waves show a sudden change in velocity. The first of these—the so-called Mohorovičić discontinuity (also known as the M discontinuity)—occurs at various depths, depending on the locality. Beneath the continents its depth ranges from 20 to 70 km; beneath the oceans it rises to a depth of 10 to 12 km below sea level. This discontinuity marks a sharp increase in the velocity of the longitudinal waves from 6.3 to 7.8 km/sec and of the transverse waves from 3.7 to 4.4 km/sec. It is now believed that the

Mohorovičić discontinuity is the lower surface of the earth's crust, which at various localities has not only a different depth but also a different internal structure. The crust will be examined in greater detail below; here we shall consider the interior zones of the earth.

Table 5. Propagation Velocities of Seismic Waves at Various Depths
(After Bullen)

Depth, km	Propagation velocities, km/sec		Shells and seismic discontinuities
	Longitudinal waves	Transverse waves	
0-35/70*	5.5-6.3*	3.2-3.7*	Crust: Mohorovičić discontinuity
35-410	7.8-9.0	4.4-5.0	Mantle: discontinuity
410-1,000	9.0-11.4	5.0-6.4	
1,000-2,700	11.4-13.6	6.4-7.3	
2,700-2,900	13.6	7.3	
2,900-4,980	8.1-10.4		Outer core: discontinuity
4,980-5,120	10.4-9.5		
5,120-6,370	11.2-11.3		Inner core

* The figures in this horizontal row apply to the continents, beneath which the average thickness of the crust is 35 km and the greatest thickness 70 km. Thus the Mohorovičić discontinuity may be encountered at various depths (see below).

The second seismic discontinuity, at which a sudden change is observed in the propagation velocities of seismic waves, occurs at a depth of 2,900 km below the earth's surface. Here the velocity of the longitudinal waves decreases sharply from 13.6 to 8.1 km/sec. This discontinuity is also the limit of propagation of the transverse waves, which do not penetrate into the deeper regions of the earth. The spherical envelope surrounding the earth between the Mohorovičić discontinuity and the discontinuity at 2,900 km is called the *mantle*. In the outer part of the mantle, between the depths of 100 and 200 km, a slight decrease in the propagation velocities of the seismic waves has been observed; this apparently corresponds to a zone in which there is some degree of "softening" of the material. The softening may be due to a combination of the rapid increase in temperature with depth within this zone and the still very small "solidifying" effect of pressure at these depths. Between the depths of 200 and 900 km, the velocity of seismic waves increases rapidly. The greater velocity is probably due to the increase in the ambient pressure; this raises the "rigidity" of the material, in conjunction with the smaller role played by temperature, which increases more slowly than pressure at these depths. On the other hand,

it has been shown (Gutenberg, 1958) that the effect of the pressure is insufficient to explain the observed increase in the propagation velocity, so some change in the composition of the material must also be assumed (see below). Between the depths of 900 and 2,900 km, the velocity of seismic waves increases extremely slowly; the increase may thus be fully explained by increasing ambient pressure in a homogeneous material.

The third seismic discontinuity has been found at depths between 4,980 km and 5,120 km. This discontinuity is not very distinct; it may be represented as a zone approximately 150 km thick within which the propagation velocity of the longitudinal waves decreases from 10.4 to 9.5 km/sec. This discontinuity is also the lower surface of the earth's outer core, which is, consequently, a spherical shell about 2,200 km thick (lying between the depths of 2,900 and 5,120 km). Since the transverse waves do not pass through the outer core and are rapidly extinguished within it, it has been concluded that the outer core is liquid. This conclusion has been supported by observations of the tides, or "solid flow," at the earth's surface.

The innermost part of the earth, finally, from the depth of 5,120 km to the earth's center, is called the *inner core*. It is characterized by an increase in the velocity of longitudinal seismic waves with depth. The state of the material in the inner core is unknown. The suggestion has been made that it is solid: this cannot be determined from the passage of seismic waves, since the longitudinal waves are propagated through any medium—solid, liquid, or gaseous—whereas the transverse waves, which pass only through solid media, have been screened out by the liquid outer core.

Between the Mohorovičić discontinuity and the earth's center, all changes in the properties of the material, as revealed by the passage of seismic waves, have been observed vertically, along the radius of the earth. There have been no reliable observations of changes in these properties in the horizontal direction, parallel to the earth's surface. All the shells, beginning with the mantle, are sharply distinguished from the earth's crust, which is characterized by great changes in structure in the horizontal direction. There is, however, some basis for supposing that small changes in the horizontal direction may be observed in the outermost layer of the mantle.

The recent extensive growth of our knowledge of the structure of the earth's crust is due chiefly to improvements in seismic methods of investigation. There has been some improvement in the methods of *general seismology*, which are based on the recording of natural earthquakes, but the main development has been in the methods of *deep seismic sounding*. These are founded on the same principles as seismic prospecting, except that the very great sensitivity of the equipment

makes it possible to explore the entire thickness of the earth's crust with waves generated by comparatively small artificial explosions. Deep seismic sounding methods have been extensively employed in the U.S.S.R., where they were developed by G. A. Gamburtsev. In somewhat different form they are also used in the U.S.A.

At the present time, deep seismic sounding and general seismic methods have been used to study the structure of the earth's crust in many regions of the Soviet Union, the U.S.A., and Europe, as well as in many parts of the oceans. Although the data obtained up to now are far from complete, they nevertheless give some general idea of the structure of the crust. Some of the regularities of this structure will be discussed briefly here.

There are two distinct types of the earth's crust: continental and oceanic. The first occurs beneath the continents and many shallow seas and beneath the continental slope, down to abyssal depths at 2,000 m; the second is encountered beneath the oceans, wherever the depth of the water exceeds 2,000 m. The continental and oceanic crusts differ above all in their thicknesses, that of the continental crust varying from 20 to 70 km, whereas the thickness of the oceanic crust is no more than 10 to 20 km, including that of the water, the mean depth of which is 5 km.

On the continents the crust has different thicknesses beneath the plains and the high mountains. In the first case, its thickness will vary from 25 to 35 km; in the second, it increases to between 40 and 80 km. Seismic observations thus confirm the existence of mountain roots, which had been suggested earlier by gravimetric data. It has also been noted that the thickness of the crust on the continents is related to the age of the folding in different regions. Its thickness is greatest, from 50 to 80 km, beneath young geosynclines, such as that of the Alpine cycle, whose development came to an end only in recent times—for example, in the Caucasus. Beneath the Hercynian folded zones, which have been transformed into epi-Hercynian platforms, the earth's crust is approximately 40 km thick. Beneath older platforms, which lie upon folded basements of Caledonian or Precambrian age, the thickness of the crust is usually no greater than 30 km.

The continental crust characteristically has two layers. In the upper layer the seismic waves travel at the velocity of 5.5 km/sec; in the lower layer their velocity is 6.5 km/sec or a little more. On the basis of a comparison of the velocities in these two layers with the elastic properties of the most common rocks in the earth's crust, the upper layer has been called the *granitic layer* and the lower the *basaltic layer*. The name "granitic layer" is, of course, arbitrary, since it includes the ancient granite-gneiss basement complex, which occurs everywhere and is primarily of Archean age, as well as the younger metamorphic rocks

and sedimentary rocks of various ages. As regards the basaltic layer, geologic data indicate that this name accurately characterizes the deeper parts of the earth's crust. The main evidence for this is the extremely homogeneous basaltic extrusives that poured out on a very large scale in different places, under the quiescent tectonic conditions of platforms, through deep fissures extending into the lowermost regions of the crust. These extrusives, and the change in velocity, indicate the presence of large amounts of basalt at the bottom of the crust. In some cases a distinct boundary (the Conrad boundary) has been encountered between the granitic and basaltic layers, at depths of 8 to 25 km. In other places, however, this boundary is apparently lacking, and one layer merges gradually into the other.

Various relationships have been observed between the thickness of these two layers and different tectonic zones. Beneath young geosynclines, for example, and particularly beneath the uplifts within them, the granitic layer is thicker than the basaltic. In general, the granitic layer is here about 30 km and the basaltic layer about 25 km thick. In the Kazbek district of the Caucasus, however, the thickness of the granitic layer reaches 37 to 46 km, whereas that of the basaltic layer is no more than 16 km. On the other hand, beneath the depressions in these young geosynclines the thickness of the granitic layer decreases sharply: in the Caucasus, under these conditions, it was found that the granitic layer is only 15 km, whereas the basaltic layer is 25 km thick. Thus the mountain roots in young geosynclines take the form of an expansion in the thickness of the granitic layer beneath the mountains. Beneath the epi-Hercynian platforms the granitic layer is somewhat thinner than the basaltic; here the thickness of the granitic layer varies from 10 to 20 km and that of the basaltic from 20 to 25 km. A much greater difference, finally, is observed beneath the ancient platforms with Precambrian folded basements, where the thickness of the granitic layer drops to 5 km and that of the basaltic layer increases to 25 or 30 km.

These figures reveal an extremely interesting progression in the structure of the earth's crust, from the young geosynclines to the ancient platforms, toward an increase in the proportion of the basaltic layer and a corresponding decrease in that of the granitic layer. Since it is well known that in geosynclines the granites are both uplifted and accumulated at the bottom of the crust in the formation of deep mountain roots, it may be assumed that two contradictory processes take place within the earth's crust: an increase in the granitic layer during the geosynclinal stage, and a decrease in its thickness during the platform stage, which continues to develop as the platform becomes older.

A study of the earth's crust beneath the ridges and valleys of the Tien Shan mountain region has shown that such "active" platforms have their own peculiar structure. Here a normal epi-Hercynian plat-

form crust has been found beneath the valleys: the granitic layer is 20 km and the basaltic layer 25 km thick. Below the upwarps, however, the basaltic layer increases to 40 km and the granitic layer decreases to 10 km in thickness. In this case, we have *basaltic* roots, which distinguish active platforms from young geosynclines. Thus mountain roots may evidently be of two kinds: granitic and basaltic.

The oceanic type of crust differs from the continental in the absence of the granitic layer. Here, beneath the water (the mean depth of which is 5 km), there is a thin layer of sediments (about 1 km thick), and immediately below this is a basaltic layer about 5 km thick. This is the structure of the earth's crust beneath the deeper parts of all the oceans. The latest data show no differences whatever in this respect between, for example, the Atlantic and Indian Oceans, on the one hand, and the Pacific Ocean, on the other.

The transition within the crust between continents and oceans has still been insufficiently studied. The information available thus far, however, indicates that at the edges of the continents the granitic layer wedges out and the Mohorovičić discontinuity rises toward the surface; in addition, a continental crust of unusual thinness has been observed beneath the edges of the continents. In California, for example, at the shore of the Pacific Ocean, the earth's crust is no more than 18 km thick, whereas a short distance to the east, below the Sierra Nevada mountains, its thickness increases to 50 km. A gradual decrease in the depth of the Mohorovičić discontinuity may be observed in approaching the North American continent from the Atlantic Ocean: beneath the open ocean the discontinuity lies at a depth of 11 km below the water's surface; at the edge of the continental shelf (150 km from the shore) it appears at the depth of 20 to 25 km; and beneath the shore line it has been found at the depth of 30 to 35 km. Sometimes the thickness of the earth's crust, in moving from the continent to the ocean, changes stepwise, by stages, so that the depth of the Mohorovičić discontinuity also changes by steps.

Beneath certain "marginal" seas intermediate structures of the earth's crust have been observed that combine elements of both the typical continental and typical oceanic crusts. Such structures have been found beneath the Caribbean Sea, the Sea of Okhotsk, the Mediterranean Sea, the southern part of the Caspian Sea, and the Gulf of Mexico. In the case of certain of these (the southern part of the Caspian Sea and the Gulf of Mexico), the crust is of the oceanic type, in the sense that it lacks a granitic layer with seismic wave velocities characteristic of granite as a rock. But here the basaltic layer is covered by a very thick (many kilometers) series of unconsolidated sedimentary rock. In other cases, such as that of the Caribbean Sea, propagation velocities of seismic waves intermediate between granitic and basaltic have been observed.

In still other cases, finally, as at the edges of the continents, a decrease in the total thickness of the crust has been found, with corresponding decreases in the separate thicknesses of its granitic and basaltic layers.

Beneath island arcs, such as the Kuril Islands or the Antilles, there is a sudden local increase in the thickness of the crust, compared with its thickness beneath the ocean, along with the appearance of the granitic layer in a narrow belt that rapidly wedges out laterally both toward the open ocean and toward the marginal sea enclosed by the island arc. Below Puerto Rico, for example, the crust is 25 km thick; as one moves from this island toward the deep trench in the ocean bottom that girdles the outer, or convex, side of the Antilles, the thickness of the crust rapidly decreases to its normal magnitude beneath the oceans. This observation shows that the formation of such deep ocean trenches, which regularly parallel the outer sides of island arcs, cannot be due to horizontal compression of the crust, as has sometimes been thought. Their formation is, on the contrary, more likely associated with expansion of the earth's crust.

CHAPTER 4

THE ORIGIN OF THE EARTH

For a true understanding of the earth's structure and history, one must know how this planet was formed and what it was like in the very earliest, or pregeologic, stage of its existence.

The question of the earth's origin cannot be separated from that of the origin of the solar system as a whole. The principal body of the solar system is the sun, and the ten bodies revolving about it are the planets. These planets move around the sun in elliptical orbits. The orbits of Mercury, Mars, and Pluto diverge considerably from that of a circle, but those of the other planets have remarkably little eccentricity. All the planets move in the same direction (counterclockwise, or from right to left if one observes them from the north) and approximately in the same plane. Only Pluto, the most distant, has an orbit inclined 17° to the plane of the ecliptic. This regularity is one piece of evidence in favor of a common origin for the planets and constitutes a starting point for the construction of a cosmogenetic hypothesis.

One very significant peculiarity of the solar system is that almost all its mass is concentrated in the sun; the planets possess only $\frac{1}{750}$ of the mass of the system. On the other hand, the angular momentum is concentrated almost entirely in the orbital movements of the planets. This inequality in the distribution of the mass and motion of the solar system constitutes the greatest obstacle to the cosmogenetic hypotheses that have thus far been proposed.

Many planets have satellites, or small bodies that revolve about them. In all, 31 satellites are known: Earth has 1 (the moon), Mars 2, Jupiter 12, Saturn 9, Uranus 5, and Neptune 2. Most of the satellites have a regular motion; that is, they rotate about the planets in the direction of the axial revolution of the planets themselves. Only 4 satellites of Jupiter, 1 satellite of Saturn, and 1 satellite of Neptune move in the direction opposite to that of their planets.

Besides the large planets and their satellites, there is a multitude of small planets, or asteroids. The orbits of most asteroids are located between those of Mars and Jupiter. About sixteen hundred asteroids are

known at the present time; of these, only a few have diameters as great as a few hundred kilometers. The diameters of most of the others are measured in kilometers or scores of kilometers. There may quite possibly be a continuous transition from small asteroids through large meteorites to cosmic dust particles. The solar system also contains numerous comets, many of which have receded to great distances from the sun. Hundreds of them have passed beyond the limits of this planetary system, which is about six billion kilometers in diameter. There is no great celestial body at a distance greater than ten thousand times the diameter of the planetary system.

Some of the physical properties of the planets are of interest. Two groups of planets can be distinguished: the planets on the scale of Earth, and the larger planets. Mars, Venus, Earth, and Mercury belong to the first group, the planets that are closer to the sun; Jupiter, Saturn, Uranus, and Neptune belong to the group of large planets. As for Pluto, in size it is closer to the Earthlike planets. Owing to the great distance between Pluto and Earth, however, practically nothing is known of its physical properties.

The inner, smaller planets are distinguished by their great density and small mass, whereas the exterior, larger planets are characterized by small density and great mass. The two groups can also be clearly distinguished by the composition of their atmospheres. Of the Earth-type planets, Earth has the densest atmosphere, consisting almost entirely of nitrogen and oxygen with only small traces of carbon dioxide, methane, hydrogen, rare gases, and so forth. Venus has an atmosphere that, so far as is known, contains only carbon dioxide; no signs of oxygen have been found. Thus, in the composition of its atmosphere, Venus differs sharply from Earth. This is the more surprising when it is considered that both planets are almost identical in size and mass. The reason for this difference evidently is the presence of life on Earth and its absence on Venus. Originally, the terrestrial atmosphere was probably like that of Venus, but the more congenial temperatures on Earth favored the appearance of living organisms. The growth of vegetation consumes carbon and releases oxygen. Earth's oxygen hence is predominately of biochemical origin.

There is a considerably greater similarity between the atmospheres of Earth and Mars. During the Martian winter white snowcaps appear on the planet, and they disappear in summer. This indicates the presence of moisture in the atmosphere of Mars. In the summer, the surface of the planet darkens in many places and takes on green hues. At the present time, it is generally considered most likely that this seasonal change in color is related to seasonal change in the distribution and state of vegetation. The atmosphere of Mars, however, has considerably less oxygen and water vapor than that of Earth: the oxygen of Mars

amounts to less than 1 per cent and water vapor to less than 5 per cent of their content in Earth's atmosphere.

Mercury apparently has no atmosphere at all. Because of its small mass and its comparatively high surface temperatures, due to its proximity to the sun, Mercury evidently did not retain its gaseous envelope, which was accordingly dissipated into surrounding space.

The major planets have entirely different gaseous envelopes. Their atmospheres have exceptionally large diameters and contain vast quantities of methane and ammonia.

Surficial processes, analogous to the geologic processes taking place on Earth, may occur on other Earth-type planets and also on the moon. The occurrence of such processes will depend not only on the solar radiation received by the planets but also on the composition and density of their atmospheres. If the composition of atmosphere differs, the surficial processes will also necessarily be different.

But exogenic processes very substantially influence the development of endogenic processes, since the latter depend not only on internal forces but also on the materials to which these forces are applied and which are produced by exogenic processes. Tectonic processes, for example, depend to a large extent on the accumulation of sediments in basins or areas of subsidence. In the same way, factors that affect the composition of the atmosphere, such as the mass of the planets or their distance from the sun (since the development or absence of life depends on the distance between the planets and the sun), indirectly influence the course of tectonic processes. There is, therefore, no reason to suppose that the structural history of all the planets was the same. Planets with different masses and distances from the sun clearly will follow different paths of development.

Kant and Laplace were the founders of cosmogony. Kant was the first to suggest that the basic laws of the solar system were not due to chance but were the result of the regular, natural development of this system; his views were later elaborated by Laplace. Thus Kant and Laplace first introduced the concept of evolution into cosmogony. From their analysis of the regularities in the solar system, they proposed hypotheses of its origin. These hypotheses had a common basis but differed in some essential details.

According to Kant's hypothesis, published in 1775, the matter that now constitutes the solar system was originally more or less uniformly disseminated throughout space and was broken down into elementary particles. In the properties of this material, Kant saw the motive force in the natural growth of the solar system. He believed that the elementary particles differed in their properties, above all in their densities. Under the influence of gravitational attraction, less dense particles were drawn into the more dense particles, and finally, where the force of

gravitational attraction was the strongest, a central body, the sun, began to form.

The result of this process was a sun surrounded by a dispersed cloud of particles, which rained down upon it. At this stage of the sun's development, Kant believed, forces of repulsion began to affect the matter in this rarefied condition (gas, smoke, or dust). Thus some of the particles that had been drawn toward the sun acquired a lateral velocity and began to rotate about it. As a result, the sun was surrounded by a huge but relatively rarefied cloud of particles, which rotated around it in different orbits and in all possible planes. The particles collided and thus changed the directions of their movements; finally, their paths gradually became more uniform. Hence, Kant thought, in the course of time, the sun was encircled by particles rotating about it in almost circular orbits and in almost a single plane. At this time particles with small relative velocities began to collide, so that, upon collision, the particles did not rebound but were united into a single mass instead. These became the germs of planets, which, by their gravitational attraction for the surrounding matter, swiftly grew into planets.

Half a century later, proceeding from the general theory postulated by Kant, Laplace proposed another hypothesis for the origin of the solar system. On the basis of W. Herschel's views of the evolution of nebulae, Laplace assumed that the sun, like the other stars, was formed from gaseous nebulae through their disintegration into separate parts and condensation under the influence of the force of universal gravitation. The primordial sun, according to Laplace, consisted of a nebula with a somewhat denser nucleus surrounded by rarefied gases, analogous to an atmosphere, which extended beyond the orbit of Neptune. Laplace further postulated that this cloud revolved about itself. Upon cooling, the cloud contracted and its rotational velocity increased, in conformity with the laws of mechanics. Simultaneously, the centrifugal forces grew; when they counterbalanced the gravitational forces on the equator of the revolving cloud, gaseous material began to separate along the equatorial plane. Particles of this material described circular orbits about the center of the nebula. Upon cooling, this material was separated from the nebula in the form of concentric rings and began to divide into individual globules. Fusion of these globules ultimately led to the formation of the planets, and the enormous central sphere became the sun.

Thus created, the planets rotated in the same direction as the cloud, in almost circular orbits and in almost a single plane: the equatorial plane of the cloud. In this way Laplace explained the regularity of the movement in the planetary system. Thus in the initial stages of growth the planets were formed in the same way as the sun, but on a smaller

scale. By the same process, the planets formed satellites about themselves. Comets, in Laplace's view, did not originate from the central mass of the solar system. He regarded comets as miniature nebulae that were drawn into the field of gravitational attraction of the solar system.

Laplace described his hypothesis very briefly and entirely in qualitative form, without any kind of mathematical treatment. Subsequently, when mathematical calculations were developed to analyze his hypothesis, very significant discrepancies were discovered, with the result that, in spite of the efforts of many great scientists to save the hypothesis through various corrections and changes, it had to be rejected.

The principal discrepancy was that Laplace's theory failed to account for the observed distribution of the mass and angular momentum in the solar system. At the time of their separation, the gaseous rings would have had to have the same rotational velocity as the nebula as a whole, and the movement of the rings would have had to be transferred to the planets that were formed from them. At the time the ring that produced the planet Neptune was separated from the nebula, the angular velocity of the nebula would have had to be the same as that of Neptune at the present time. Further, it has been calculated that, with its shrinkage, the rotational velocity of the nebula should have gradually increased. In addition, one may determine what rotational velocity would be possessed at the present time by the relict core of the nebula (that is, by the sun) if events had occurred as suggested by this hypothesis. It turns out, however, that the observed facts do not agree at all with the theory. The sun rotates very slowly, many times more slowly than it should according to this hypothesis. A second difficulty is that the heated, gaseous material separating from the equator of the revolving nebula would have been dispersed in space and not consolidated into planets. The direction of the planets' axial revolution is also in contradiction to the hypothesis. Thus the attempts to overcome these obstacles did not meet with success, and after a prolonged period, lasting the entire nineteenth century, the Laplace hypothesis was finally abandoned.

Thereafter various new ideas appeared. For example, the hypothesis of Moulton and Chamberlin aroused considerable interest when it was introduced at the beginning of the twentieth century. According to this theory, the solar system was formed when the sun was closely approached by another star. As a result of the tides created in the sun's surface by the gravitational attraction of the other star, a multitude of gaseous streams of various sizes were drawn from the sun's surface. Many of these were dispersed to form a cloud around the sun; others, however, began to draw together into masses of different sizes. These gaseous masses cooled very quickly and passed into the solid phase. Consequently, the space around the sun was filled with solid bodies of varying but small sizes, down to the finest dust particles. These bodies,

moving in very different orbits and directions, collided with one another and, under the forces of mutual attraction, coalesced. The larger ones, having greater mass, acted as centers of condensation. Small bodies fell into the larger ones, just as meteorites fall into the earth (but much more frequently, as a continuous rain of solid particles), and from these increments the planets originated. The collisions between particles altered their orbits, which began to approximate a circle, and the bodies began to move in a single direction.

This theory, as will be seen when it is compared with O. Yu. Schmidt's modern cosmogenetic theory, to be described presently, was an important step forward. Like Laplace's hypothesis, however, it cannot explain the unequal distribution of mass and momentum in the solar system, apart from many other important phenomena.

At this point, Jeans's hypothesis took the lead and remained in favor for some twenty years. This theory also holds that the solar system originated when a star passed very close to the sun. The intruding star, by the force of its gravitational attraction, tore away the material from the sun's surface into a long stream. When the star receded, one part of the stream of gas fell back into the sun; another part was dispersed and formed a cloud around the sun, while still a third, which was larger and denser, separated into large globules. By condensing and cooling, these developed into the planets. The planets then began to rotate in elongated orbits about the sun, at times approaching very close and then moving great distances away. By cooling rapidly, some of the gaseous bodies became liquid. When the liquefied planetary bodies first passed close to the sun, the tides induced in them by the latter tore off streams of matter from their surfaces, thereby forming the satellites. Because of the resistance encountered by the planets on their paths through the dispersed gases, their orbits gradually changed and became almost circular.

Jeans's hypothesis also had a number of important physical and mathematical shortcomings. Like the previous theories, it was unable to reconcile the distribution of the mass with that of the angular momentum in the solar system. An essential methodologic defect of this theory is its assumption of the exclusiveness of the solar system: the passage of two stars at such a small distance from each other, as the Jeans hypothesis requires, is so unlikely an event that planetary systems, if they did originate in this way, would be extreme rarities in the universe.

Jeans's hypothesis was rejected in the late nineteen thirties and early forties, after the Soviet scientist N. N. Pariyskiy had raised decisive objections against it.

Since that time, scientists have proposed a great number of various cosmogenetic hypotheses, most of which have been impromptu and not carefully thought out. Thus they have vanished as quickly as they ap-

peared. Because of these failures, a number of important scientists have concluded that it is not possible at the present time to formulate a satisfactory theory for the origin of the solar system, and even that the problem is outside the scope of scientific analysis in general.

In spite of this negative outlook, however, considerable success has been achieved in recent years in the development of cosmogenetic theories. This success is due not only to the work of individual scientists but to the efforts of the entire scientific community. At this point, one may mention the work of Academician V. G. Fesenkov, who has considered a number of possible ways in which the solar system may have evolved.

The most highly developed theory of the origin of the solar system is that of Academician O. Yu. Shmidt and his colleagues. This was first advanced in 1944 and has since been revised and improved (Shmidt, 1950; *Transactions of the First Conference on Problems of Cosmogony, April 16-19, 1951*). To determine the state of the matter of which the solar system was created immediately before the formation of the planets, Shmidt made a detailed analysis of the fundamental laws of the solar system. His approach was the same as Kant's and Laplace's, but it was raised to the level of the science and philosophy of the present time. There is no room for a detailed discussion of this analysis, beyond noting only a few of its characteristic features.

One of the most important regularities indicating the former state of the material from which the planets were formed is the circular nature of the planetary orbits. This shape of the orbits suggests that the planets were formed through a series of collisions of a multitude of small particles, orbiting in a huge swarm about the sun. According to the laws of mechanics, the motion of isolated particles in circular orbits constitutes an extraordinary event of insignificantly minute probability. As a rule, the individual particles, in the early stages, must have rotated about the sun in elliptical orbits in various directions. Only after numerous collisions and changes in velocity and after the combination of tiny particles into larger ones—embryonic planets—were they able to attain uniform, symmetrical orbits of almost circular shape.

In addition, there are physicochemical data indicating that the planets were formed from such "dustlike" material. The composition of the planets themselves and of their atmospheres suggests this conclusion. There are no chemical elements on the earth that could not have been contained in these solid dust particles. For example, the sun and the cosmos in general have almost equal quantities of oxygen and nitrogen; but in the earth's crust and its atmosphere there is ten thousand times less inert nitrogen than chemically active oxygen.

A theoretical study of the processes in the dust cloud, which might have contained a trace of gas as well, shows not only that coalescence of

the particles into planets is possible but also that the planets formed in such a manner would have to revolve on their axes in the same direction as they rotate about the sun. None of the previous theories was able to explain this fact.

Thus immediately before the formation of the planets, the solar system consisted of the sun surrounded by a swarm of dust-sized particles rotating about it in elliptical orbits. This swarm had a high angular momentum: the number of particles moving in the common direction was much greater than the number of particles moving in various other directions. The planets, comets, and meteors were formed by the fusion of particles. Gas molecules present in the primordial cloud froze and settled upon the solid particles.

Shmidt's theory explains the more important properties of the solar system. As we have seen previously, the circular character of the planetary orbits is due to the averaging of individual elliptical orbits. This is confirmed by the fact that the orbits of the very large planets most closely approach a circle, whereas those of asteroids, and especially of comets, have much more eccentricity; in many cases the eccentricity is very great. The larger planets rotate in planes close to the fundamental plane of the entire planetary system, and the orbital planes of asteroids and comets are considerably inclined to the ecliptic.

After studying the evolution of the swarm of particles captured by the sun, O. Yu. Shmidt showed that the larger bodies of the swarm could easily become the embryos of future planets. The paths of both small and large particles would intersect, and the small particles would fall into the larger ones and combine with them, gradually forming planets at the expense of the particle swarm. Investigations by L. E. Gurevich and A. I. Lebedinskiy have shown that such a condensation of the particles into planets would have to take place even if there were no very large particles in the swarm to act as centers of condensation. It has been established that the dissipation as heat of part of the energy produced by collisions must help to concentrate the swarm, increasing the frequency of collisions and accelerating the condensation. At the same time, it must decrease the velocity of particles relative to each other, so that after a certain critical average relative velocity and density have been reached, the swarm could not retain its previous form. Under the influence of gravitational attraction, it would rapidly condense and form clots. At first, these clots would be on the order of asteroids in size, but ultimately they would have to develop into a small number of very large bodies: planets.

One important factor in the evolution of the swarm was the pressure of light from the sun. This had the effect of hindering the movement of particles relatively close to the sun, causing many of them to be drawn into it. This depleted the matter in the inner part of the swarm and

explains the separation of the planets into two groups of small planets close to the sun and large planets further away. The small planet Pluto was formed at such a distance that the swarm had dwindled to practically nothing. Meanwhile, the particles that fell into the sun transferred their own momentum to it, causing the sun in time to revolve about an axis approximately perpendicular to the plane of the planetary system.

The satellites, according to Shmidt's theory, were formed during the condensation of the swarm, as an accessory result of the formation of the planets themselves. The theory also explains why some satellites have a regular movement in the same direction as that of the planets and others move in an opposite direction.

Shmidt's theory won recognition in the scientific circles of the Soviet Union. The problem inevitably arises, however, of the origin of the cloud of dust particles, or "planetesimals," about the sun. In Shmidt's theory, the sun, in moving through the galaxy, encountered a cloud of minute particles, of which there are many in the galaxy, and captured it by gravitational attraction. The possibility of such a capture was mathematically proved by Shmidt and Khil'mi in 1947 and 1948. At the present time, such capture is the only process capable of explaining the distribution of angular momentum in the solar system.

Nevertheless, the supposition that the sun appropriated a cloud of galactic dust, originating independently of the sun, aroused many disputes and objections. This idea has been seriously questioned. For example, it was argued that if the velocity of the sun relative to the cloud was great, then the probability of capture was very small. It is more likely that the presence of the cloud around the sun was in some way related to the formation of the sun itself or of a group of bodies that included the sun. The concept of capture in this scheme requires further elaboration.

Shmidt's theory is particularly valuable in connection with the subject of this book, since it has very important geologic and geophysical consequences and also provides a possible connection between the pregeologic and geologic stages in this planet's history, which was found to be lacking in the other theories. More will be said about this in the following section.

THE PREGEOLOGIC STAGES OF THE EARTH'S HISTORY

For a long time geologists, influenced by Laplace's hypothesis, supposed that the earth was originally molten. Thus the pregeologic stages in the development of the planet were looked upon as a process of gradual cooling and solidification. The duration of the solidification and its mechanism were specially investigated. The original molten con-

dition of the earth was directly related to its further development, in accordance with the contraction hypothesis. We have seen, however, that the contraction hypothesis in its classical form must be rejected. Modern ideas on the internal structure and great solidity of the earth do not require us to assume that it was once in a molten state.

According to Shmidt's theory, the earth was never molten. Like the other planets, it was formed from cold, solid particles. At first, when the swarm of particles was still dense, the process of their collection into the planet went very quickly, but it gradually slowed down as the material in the swarm was exhausted. Particles of different chemical composition coalesced in a random manner, so that the earth, in its early stages, had a uniform composition, both light and heavy particles being contained in it without any order.

The distribution of matter observed in the surficial layers of the earth indicates that here the original uniform composition was later effaced and that granitic and basaltic layers have been formed. This separation of the material in the upper region of the earth is undoubtedly connected with gravitational differentiation, expressed in the rising of the lighter material (granitic) and the sinking of the heavier material (basaltic). This is concretely manifested in the progressive enrichment of the earth's crust by granitic intrusions. Differentiation in the crust has thus taken place throughout geologic history and is continuing today. Quite young granitic intrusions have been observed in geosynclines. Ye. N. Lyustikh (1948a) thinks that the gravitational differentiation process is entirely capable of providing the energy required for tectonic movements; this will be seen below. O. Yu. Shmidt believes that the differentiation of matter in the earth was facilitated by heating in the earliest stages of the planet's existence, as a result of decay of the radioactive elements contained in the protoplanetary substance. Such heating would promote mobility of the substance and favor rapid ascension of the lighter and sinking of the heavier materials. Shmidt postulated that the heating of the earth is still taking place but is probably approaching a maximum, after which the earth will slowly cool because of the decreased amount of radioactive elements. We have seen that, according to the latest data from Ye. A. Lyubimova, the interior of the earth continues to be heated while the outer layers are cooling off. Shmidt's idea that the earth has developed its own internal heat during its history is one of exceptional interest and deserves further study by geologists, geochemists, and geophysicists. It is reasonable to suppose that the intensive heating (up to the melting point) that occurs in places where radioactive elements are accumulated has had, and still has, considerably greater significance than overall slow heating in the development of the earth's internal structure. Such local and temporary melting of rock, occurring first in one place and then in another and in the course

of time encompassing vast regions of the planet, removes all the objections against a primordial cold earth that have been raised by geochemists. Geochemists believe that the structure and composition of many rocks, as well as the distribution of elements in them, indicate that the rocks must have gone through a period of fusion. Such a stage may certainly have occurred, but not in the primordial earth. As S. D. Chetverikov has pointed out (*Transactions of the First Conference on Problems of Cosmogony, April 16-19, 1951*), periodic melting affords a better explanation of the composition of igneous rocks.

Hence magmatic processes, intrusive as well as extrusive, have no connection with any remnants of originally molten material in the earth's interior, but they arise rather from secondary, local, and intermittent melting due to the concentration of radioactive elements. Such an accumulation is necessary to explain the differentiation of the material in the earth. This melting or fusion accelerated the differentiation, aiding in rapid separation of the materials in the outer layers and facilitating the ascent of granite in a molten state to form the crust.

In the course of their ascent, small particles coalesced into larger bodies or streams. Thus differentiation led to the formation of slowly rising and sinking streams of matter, which were probably more dispersed at depths and more concentrated toward the surface. We shall see later that these subcrustal currents may be related to oscillatory movements of the earth's crust, which are an important type of tectonic movement.

The particles, some rising and some sinking, passed through regions of the earth with different pressures, temperatures, and other physical conditions. In its journey, the matter could pass from one state of aggregation to another and change its molecular structure. The earth's substance can exist in a crystalline state only close to the surface, in a region of comparatively low pressure. This upper, crystalline shell is the earth's crust, the outermost part of which consists primarily of granite. The crust was formed gradually by the differentiation and concentration of acidic material close to the surface. The granitic layer was formed essentially before the end of the Archean era, only an insignificant amount being added later. In view of its crystalline state, the properties of the crust must differ very markedly from those of the deeper, amorphous zones of the earth. This difference plays an important role in tectonic processes.

The deep-seated subcrustal matter, existing in an amorphous state under conditions of great pressure, consists of volatile gases and liquefied components. This matter has the ability to react vigorously to conditions in the upper regions of the earth. Even a small rise in temperature will cause partial rarefaction of the material, thus increasing its volume and mobility. The temperatures at which matter freezes and

melts are approximately the same, so that as the temperature fluctuates about the freezing point, the state of the matter continuously changes from solid to liquid and back again.

Crystalline materials behave differently from amorphous substances. Upon freezing, which accompanies crystallization, there is a loss of volatile components, particularly water; this results in a significant rise in the melting point (Goranson, 1932). It has been observed that basaltic magma still saturated with volatiles remains fluid in the crater of Kilauea at 600 C. But after freezing and crystallization, the basalt will begin to melt only at 1200°. Goranson studied the effect of the volatile components on the melting point of granites and found that a granitic magma with a water content of 9 per cent (which Goranson believed to be the water content of granitic magma at depth) will freeze at 720°; but after the magma has crystallized and lost its volatiles, it will remelt only after the temperature rises above 1000°. Thus the process of freezing and crystallization is not entirely reversible. After the material has solidified and passed over to a crystalline state, it cannot be liquefied by raising its temperature to the original melting point. In order to remelt, considerably higher temperatures are required. Moreover, after remelting, its composition will be different, since it has lost its volatile constituents.

Consequently, the material of the earth's outer crystalline shell would be considerably less mobile and more inert than the amorphous material at greater depths. Whereas a small change in temperature must produce a corresponding melting or freezing at depth, near the surface the same change in temperature will not cause any noticeable change in the state of the material. This remarkable inertness appears to be among the primary properties that distinguish the crystalline crust from the interior of the planet.

There is also a considerably greater change in volume when a substance crystallizes than when an amorphous material solidifies. The loss of volatiles has a bearing on this change in volume. As a result, greater binding stresses must develop in the crystalline crust and compress the interior of the earth as the tight cover of a football compresses the air inside it. This is also facilitated by the supposed cooling of the earth's outer layers. Just as the crystalline shell compresses the material in the interior of the planet, so does this material, reacting to continuous changes in temperature and pressure, change its volume and exert a corresponding mechanical counteraction on the upper crust. The latter resists these effects, placing a strain on the earth's contents, and tries to keep the earth's volume at a minimum.

The present author believes that this contest between the more mobile material within the earth and its restraining cover is an important factor in the structural development of the planet in the geologic stage of its

history. Having established this difference between the deeper regions and the outer shell, we can now define more precisely the meaning of the "earth's crust." Previously, it was thought that the interior of the earth was still molten and that the crust was a hard shell enveloping the liquid core. After it had been established that the entire earth is solid, the old conceptions could no longer be retained and many researchers came to believe that all ideas on the crust in general were obsolete. Following Clark, they defined the crust as that part of the earth which was exposed to processes of erosion and accessible to direct observation. Clark thought that the maximum thickness of this part of the earth amounted, in American units, to 10 miles (16 km). It need hardly be said that such a concept of the earth's crust is arbitrary and artificial.

On the basis of the previous discussion of amorphous and crystalline materials, however, the old conception of the crust may be reexamined and given new meaning. The crust is quite properly considered to be the outer crystalline shell of the planet, which constrains the deeper-lying amorphous material and binds it together, resisting its movement and possessing unique mechanical properties. Finally, the lower boundary of the crust is, in this view, the Mohorovičić discontinuity.

CHEMICAL COMPOSITION OF THE EARTH'S INTERIOR

The following distribution of material has been observed in the upper part of the continental crust, to depths of several score kilometers: a layer of granite at the top, and a layer of basalt below. Hence there is a change in the composition of matter with depth, in the sense of a reduction in silica, sodium, and potassium oxides and an increase in iron, magnesium, and calcium oxides. The density also increases with depth.

Can this data be extrapolated to determine the composition at much greater depths? Can the increase in density toward the center of the earth be regarded as the result of changes in composition—a gradual relative increase in the metal content (principally iron) and a decrease in silicon and oxygen? During this generation there has been a basic change in the approach to these problems. In the nineteenth century, it was generally believed that the high density of the earth's core could not be explained purely by the pressure exerted by the overlying mass. It was firmly believed that rocks such as those of the surface could not, as a result of pressures in the center of the earth, increase in density from about 3 to 10. It was therefore concluded that the material of the earth's core was, by virtue of its composition, intrinsically more dense and quite different from the material of the outer zones of the earth.

Wiechert was the first, in 1910, to suggest a model earth consisting of two zones. He believed that the earth contains a metallic core and a

rocky shell. The boundary between them he placed at the depth of 1,600 km, on the basis of seismic data that, even for those days, were inaccurate and insufficient.

V. Goldschmidt's scheme, developed in 1922, enjoyed great popularity. Goldschmidt based his hypothesis on a large amount of data. In the first place, he believed that the chemical composition of the earth as a whole had to be close to that of the mean composition of meteorites, since meteorites originated in the disintegration of a planet, or so it was believed at the time. In making this assumption, Goldschmidt was following the Russian scientist P. N. Chirvinskiy, who in 1919 first determined the average composition of the earth by means of meteorites. The existence of two main groups of meteorites—iron and stony—was Goldschmidt's reason for supposing that these two groups correspond to the two main layers of the planet, which must also exist within the earth. Goldschmidt calculated new data on the depths of seismic wave discontinuities and announced a theory that these boundaries corresponded to sharp changes in composition. He took into account the distribution of densities along the earth's radius. Finally, on the basis of the idea that the earth was originally in a hot, molten state, he drew an analogy between the processes occurring in a molten mass of planetary dimensions and those in a blast furnace. Upon melting in a blast furnace, the ore separates into three layers: an upper and lightest layer of silicate slag, a heavier layer of oxides and metallic sulfides, and the very heavy metals on the bottom.

Goldschmidt suggested that in the molten earth there was just such a process of separation of the material by density, causing the heaviest material to be concentrated near the center and the lighter material closer to the surface. This differentiation, in Goldschmidt's opinion, occurred in the primordial, "hot" stage of the earth's history.

On the basis of all these considerations, Goldschmidt produced the scheme of the earth's structure that is illustrated in Table 6. This was

Table 6. The Geospheres (After V. Goldschmidt)

Depth, km	Geosphere	General composition of the geosphere
0-60	Silicate	Silicates: the common rock-formers; predominantly granite above (to 20 km) and basalt below
60-1,200	Eclogite (ultrabasaltic)	Basic silicates, principally olivine
1,200-2,900	Sulfoxide (ore)	Metallic sulfides, mostly iron
2,900-6,370	Central core	Nickel-iron

more or less modified by later investigators, who did not, however, alter it in principle.

According to this scheme, the core of the earth has a composition

similar to that of the iron meteorites. The latter consist, on the average, of 91 per cent (by weight) iron, 8 per cent nickel, and 1 per cent cobalt and phosphorus. The composition of the sulfoxide and eclogite shells, located between the core and the crust, as suggested by A. Ye. Fersman, is presented in Table 7.

Table 7. Chemical Composition of the Eclogite and Sulfoxide Geospheres (After A. Ye. Fersman)

Element	Eclogite layer	Sulfoxide layer
O	42.05	29.10
Si	23.00	13.84
Fe	13.50	37.03
Mg	10.91	13.62
Ca	5.01	0.86
Al	3.26	0.77
S	0.54	0.92
Na	0.50	
Ni	0.33	2.91
K	0.22	

In recent years this scheme of the earth's structure has been subjected to criticism. The Russian scientist V. N. Lodochnikov first suggested that the seismic discontinuities and changes in density with increasing depth were not determined so much by changes in chemical composition as by the transition of this material from one state of aggregation to another under the influence of pressure (Lodochnikov, 1939). "In my opinion," wrote Lodochnikov,

There is absolutely no reason to reject the hypothesis of physical transmutations in zones of the earth's interior (Legendre-Laplace) in favor of a theory of chemical transmutations (Wiechert, his school, and the great majority of geologists, geochemists and geophysicists). Actually, Bridgman's experiments and several examples cited above of specific gravities of chemically different substances found on the earth's surface, as well as small changes in specific gravity at shallow depths (which Wiechert himself has admitted) along with the purely physical changes in the layers with depth all the way down to the core, all favor the hypothesis of zonal change in the physical state.

Lodochnikov cited density differences found in varieties of the same material (white and black phosphorus, yellow and gray arsenic, white and gray tin) even under normal pressure. He also pointed out that the sudden change in density at the boundary of the core, as established by the change in the velocity of seismic waves, is difficult to explain by a change in the composition, since differentiation would produce a gradual, and not a sudden, change in composition. In this connection, Lodochnikov wrote:

In defense of my hypothesis, it may be said that such a discrete change in density is a simple and necessary consequence of the primary assumption of purely physical causes for the changes in the observed velocity. Although the change in density due to loading is gradual, it nevertheless reaches a certain point of transition at which the material passes from one state to another; regardless of the kind of transition—whether from one crystalline variety to another, or from a crystalline state to an amorphous state, or even if the change is from a known state of aggregation to an unknown new state all such changes must occur by discrete “jumps,” just as we have observed in the case of changes in various states of aggregation on the earth’s surface.

Since 1930, a number of foreign authors have come out against Goldschmidt’s theory (Gesztzi, 1937; Kuhn and Rittmann, 1941). Their principal objection is that the analogy between the planet, even in the molten state, and a blast furnace is entirely unsuitable. In a blast furnace the viscosity of the melt is such that a rapid and complete separation of the material by density is entirely possible. In the interior of the earth, owing to the enormous pressure, the viscosity of matter is so great that the movement of matter necessary for differentiation is, for all practical purposes, almost impossible. Differentiation would have had to take place so slowly that the entire age of the earth would not be time enough for it to achieve any discernible results. Differentiation, on a large scale can therefore develop only in the parts of the earth relatively close to the surface, where the pressure and viscosity are very small and where, in addition, the force of gravity, which causes the differentiation, is still great, since it is well known that its magnitude diminishes from depths of about one thousand kilometers to the center of the earth. Therefore the core cannot be composed of an accumulation of heavy material that had previously been uniformly distributed throughout the entire earth. The core must have a composition close to the primordial composition of the earth as a whole. In 1949, Ramsey, elaborating these views, proposed the theory that the core of the earth consists of the same silicate material that makes up the crust. But this material, under great pressure, has been transformed into a “metallic” state: the atoms have lost part of their electrons and consequently have greatly increased in density (Ramsey, 1950).

Very recently the Soviet scientist B. Yu. Levin (1953) made a detailed study of the earth’s composition, from the standpoint of the most recent information on the origin of the planets. Levin pointed out that, in accordance with the above-described theory of O. Yu. Shmidt on the origin of the planets, meteorites cannot be considered fragments of what was once an enormous planet. They represent fragments of a multitude of asteroidal bodies and have a complicated history, in the course of which they were repeatedly both fused together and separated. On the other hand, there is the very significant fact that meteorites were

formed from the same protoplanetary cloud of cold particles from which the earth was created, and also from the same part of this cloud. In view of these circumstances, we may, as before, base the determination of the earth's average composition on the average composition of meteorites.

In determining the average composition of meteorites, however, it is difficult to estimate correctly the proportion of their silicate and metallic parts. In many calculations, the proportion of iron meteorites in relation to stony meteorites has been exaggerated. Goldschmidt made this error, which was only partly corrected by A. Ye. Fersman. According to B. Yu. Levin's calculations, the ratio of metallic to silicate meteorites is 1:6; moreover, both contain 5 per cent troilite (a sulfide compound). Hence the average composition of meteorites and of the earth as well must be as shown in Table 8.

*Table 8. The Average Composition of Meteorites and the Earth
(After B. Yu. Levin)*

<i>Element</i>	<i>Per cent</i>
Oxygen.....	33.4
Iron.....	28.1
Silicon.....	16.7
Magnesium.....	12.9
Nickel....	1.3
Sulfur.....	3.2
Calcium.....	1.6
Aluminum.....	1.4
Others.....	1.4
Total.....	100.0

According to Table 8, neither metallic iron nor all the other forms of iron, including that in iron silicate compounds, are sufficient to account for a core of the dimensions observed.

B. Yu. Levin has stressed the fact that the rate of gravitational differentiation in the deep-seated layers of the earth would be insignificant even if the planet were in a molten state, so it would be even more insignificant in the light of Schmidt's theory of the "cold" origin of the earth. This limited gravitational differentiation requires us to reject Goldschmidt's views. In the deep interior of the earth, the material could not be noticeably affected by such differentiation and would have retained its initial composition: the average composition of the earth. This observation, however, does not apply to the outermost layers of the earth—the crystalline shell and the crust, where differentiation not only could take place (the more easily, the closer it was to the surface) but doubtless played a manifest part in tectonic processes. B. Yu. Levin accepts Ramsey's hypothesis of the metallic state of material in the

core resulting from the great pressures, and he attributes to it the core's physical properties: its density, solidity, and elasticity.

Certain peculiarities of the distribution of elements in the sun compared with the planets have previously been explained by differences in the mass of the individual planets and by the dissipation of light elements when the planets were hot. These peculiarities may be considerably better explained by differences in the physicochemical properties of the matter and by different conditions in the different parts of the protoplanetary cloud.

Thus the most recent data favor the hypothesis of a homogeneous core of silicate composition but existing in a unique and dense "metallized" state. This conclusion, however, cannot be applied to the surficial parts of the earth—to the crystalline outer shell and the crust. Differentiation has undoubtedly occurred in the latter, and most intensively close to the surface. Therefore the changes in composition with depth that have been firmly established in the case of the crust—meaning an increase in the relative contents of iron, magnesium, and calcium at the expense of oxygen, silicon, sodium, and potassium—may with confidence be attributed to the entire crystalline shell also. But it should be kept in mind that as the depth increases, such changes in composition occur more slowly and probably amount to nothing in the deeper parts of the crystalline outer shell.

In this connection, the works of V. A. Magnitskiy are of great interest. On the basis of physical theories of the structure of matter, he has attempted to solve certain problems of the chemical composition of the crystalline shell by calculating the modulus of compression and density (Magnitskiy, 1952). He suggested that magnesium and iron silicates and oxides must be the most important components of the shell. The mineral best corresponding to this composition is olivine. Although reliable calculations are difficult and do not yield consistent results, it is nevertheless clear that the composition of the crystalline shell must change with depth. The uppermost part of the shell may be predominantly forsterite (Mg_2SiO_4). At depths of a few hundred kilometers, the composition changes because of the breakdown of the forsterite into silica and magnesium oxide. The silica rises to the surface, and the magnesium oxide accumulates in the shell. It is also possible that at depths of a few hundred kilometers there is an increased content of iron oxide.

From the preceding remarks it is evident that the problem of the chemical composition of the earth's interior is still far from solution. Goldschmidt's oversimplified picture, which does not meet the standards of the present level of knowledge, must be rejected. New hypotheses of the composition and state of the matter within the earth, however, have only begun to appear. These are now being elaborated as the

only path to the solution of this very difficult problem. There is no doubt that, thanks to the most recent accomplishments in the field of the physical composition of matter, substantial achievements in this area will be made in the near future. This will require solving problems not only of the chemical composition of matter but also of its state and physicochemical properties. It may be assumed that the crystalline state and complex molecular compounds are characteristic of matter only at the very surface of the earth. At greater depth, evidently, there begins to be a transition to the amorphous state and to simpler compounds. At extremely great depths the matter must again change to a monatomic structure, and finally to a "metallized" state with deformed atoms.

INTERNAL ENERGY OF THE EARTH

Henceforth the discussion will deal predominantly with tectonic movements. The displacements of material that cause changes in the earth's structure are associated with the expenditure of energy. This raises the question of the source of this energy. The following are among the principal types of the earth's energy:

1. Gravitational energy (the potential energy of the earth's mass)
2. Rotational energy
3. Chemical energy, or energy of crystallization
4. Thermal energy (including radioactive energy)

Other forms of energy, such as magnetic and electric, are evidently not of great importance in tectonic processes.

Gravitational energy is manifested in the falling or subsidence of any mass. It is also manifested in the redistribution of masses within the earth, when the denser material sinks and the lighter material rises. This potential energy has repeatedly been indicated as the cause of tectonic movements.

According to the contraction hypothesis, all tectonic movements are produced by gravitational forces, since the compression of the crust is caused by settling under the influence of gravity. Although the contractional hypothesis in its original form must be rejected, the potential energy of the earth's mass has an important role in geotectogenesis, according to the latest concepts of geotectonics. There is convincing evidence in geologic and geophysical data that the main process at depth of fundamental significance in tectonics is the differentiation by density of material in the crystalline shell, with the denser substances sinking deeper and the lighter "floating" to the surface. Ye. N. Lyustikh calculated that the potential gravitational energy resulting from this would be entirely sufficient to produce geotectonic processes (Lyustikh,

1948b). According to his calculations, about 10^{28} ergs of energy have been liberated through the sinking of denser material and the rising of less dense material since the earth has been in existence. This amount of energy is fully adequate to account for all the tectonic transformations that have occurred on the earth. Lyustikh, however, believed that differentiation also occurs in the core.

Gravitational energy is manifested in another way as well: it produces tidal forces, to which many researchers attach great significance. They suggest, for example, that tidal forces may cause continental drift. This viewpoint cannot be supported, but it is not impossible that these forces, by inducing "tidal" waves in the lithosphere, contribute to the production of tectonic processes such as fracturing in rocks. At the earth's surface the force of gravity plays a leading part in the formation of folds.

The earth's rotational energy is related to centrifugal forces, which must have some influence on tectonics. But the extent and consequences of this influence are unknown. Some tectonists impute great importance to rotational energy, suggesting that centrifugal forces together with tides in the lithosphere cause the movement of continents. This view has recently been disproved. The role played by rotational energy in tectonics is probably very small.

Chemical energy and the energy of crystallization can be converted to heat energy, which causes the development of various mechanical processes. There is no definite information at all on the amount of this energy. In some cases, however, chemical processes in the earth's interior can lead to considerable local increases in temperature, and even to the melting of solid material. Nevertheless, in the overall energy balance of the earth, chemical energy and the energy of crystallization probably play secondary roles.

Heat energy undoubtedly figures very prominently in tectonic processes. A tremendous flow of heat energy is continuously received from the sun; this produces exogenic geologic processes, as well as the geochemical and biochemical phenomena manifested on the planet's surface. Solar heat is also important in the tectonic phenomena with which this book is concerned. Surficial processes, manifested in the disintegration of rocks, in the movement of particles, and in the deposition of sediments, thus redistribute the mass of the crust and, consequently, change the material on which tectonic forces operate. Changes in the material, even if the forces remain the same, must inevitably lead to changes in the character of tectonic movements. If no exogenic geologic processes took place on the earth's surface, the tectonic transformations of the crust would ultimately be different even if all the endogenic forces and processes were to remain as they are now. One reason for the sharp differences in surface forms between the earth and the moon

is probably the moon's lack of an atmosphere; thus there is an almost complete absence of exogenic processes on the moon. But in this case the effect of solar heat is not a determining factor. Solar radiation, through surficial processes, affects the result of tectonic forces, but the radiation itself does not produce these forces. Any other view will not withstand criticism.

From time to time, the suggestion is made in the literature that fluctuations in the amount of solar radiation are the cause of the periodicity in geotectonic processes. This involves the assumption that solar radiation can strongly influence the temperatures in the depths of the earth. This view, however, is incorrect: observations indicate that solar heat penetrates only a short distance into the earth. Moreover, excessively great changes in the sun's temperature would be required to produce perceptible variations in the temperature beneath the earth's crust. Internal heat originating in the depths of the planet itself is of far greater importance.

At the present time there is no doubt that almost all the heat emitted from the earth is of radioactive origin. The most recent calculations indicate that, because of the very slow diffusion of heat within the earth, the planet over a long period of time must become warmer, despite the continuous decrease in the amount of radioactive elements. Radioactive heat is thus an energy factor of planetary scale.

In summary, then, gravitation and radioactive heat are apparently the principal forms of the earth's internal energy that determine tectonic processes.

AGE OF THE EARTH

The preceding section mentioned the concept of the length of geologic time and its significance for the science of geotectonics. To comprehend the nature and mechanism of tectonic movements, one must know the rate at which they take place, their periodicity, and so forth. This requires knowing not only the age of the earth but also a scale of absolute geologic time by which the absolute duration of each stratigraphic subdivision can be established. Investigators have repeatedly and by various methods attempted to determine the earth's age and to produce an absolute time scale. Only with the introduction of the radioactive method for determining the absolute age of minerals and rocks, however, did this problem find a really scientific solution.

The radioactive method is based on the fact that radioactive elements decay and are transmuted to other elements at a rate that is practically constant under the conditions of the earth. By determining the rate of radioactive decay from the ratio observed in the minerals between the remaining radioactive material (U, Th) and its decay products (Pb^{206} ,

Pb²⁰⁸, He), one may accurately calculate the absolute age of the minerals. This calculation also fixes the time at which the mineral was formed as a solid body (Starik, 1938).

Actually, there are many technical difficulties in this method that diminish the number of cases in which it can be applied. Nevertheless, with the aid of this method, a fairly accurate scale of geologic time has been worked out, in which the durations of the individual geologic periods in absolute years are established. The durations of the shorter stratigraphic subdivisions have still not been accurately determined.

Table 9. Absolute Geologic Time Scale*

Periods	Absolute ages (10 ⁶ yr)		Duration of periods and eras (10 ⁶ yr)
	Beginning of period	End of period	
Quaternary.....	0.5	0	0.5
Neogene.....	30	0.5	30
Paleogene.....	70	30	40
Cretaceous.....	110	70	40
Jurassic.....	150	110	40
Triassic.....	185	150	35
Permian.....	225	185	40
Carboniferous.....	275	225	50
Devonian.....	310	275	35
Upper Silurian.....	345	310	35
Lower Silurian.....	390	345	45
Cambrian.....	455	390	65

* These data were compiled on the basis of information available in 1954, and text discussion is consistent with them. In discussing the periodicity of tectonic processes, it is stated they occur in cycles of approximately 150 million years; in the light of recent age data, this should be extended to about 200 million years. V. V. B.

Table 9 shows the most probable lengths of the periods and of certain eras, beginning with the Cambrian. It should be kept in mind that these figures contain a possible error of 10 to 15 per cent. Table 9 reveals an extraordinarily interesting fact: the different periods differ very little in duration. The average length of each period is about forty million years. The Paleogene and the Neogene are to be considered separate periods; this view has recently been supported by many stratigraphers. In addition, the Cambrian and Silurian periods are anomalous. Their great spans of time evidently indicate that they consist of the equivalent of two of the other periods. The division of the Silurian period into the Ordovician and Gotlandian (or Silurian proper) has now been adopted by the majority of geologists.

The nearly uniform time span of all the periods may be explained as

follows: the boundaries between periods correspond to times of significant change in the totality of the world's fauna and flora. But these changes to a considerable degree reflect large-scale tectonic movements, which produce redistributions of seas and land, changes in topography and climate, and so forth. Evidently, these movements are characterized by some degree of periodicity.

If 450 million years have passed since the beginning of the Paleozoic, what is the age of the planet? The oldest minerals on the earth's surface that have been studied are 1.6 billion years old [written in 1954]. But the age of the earth is undoubtedly much greater. According to I. Ye. Starik's calculations, the general relationship between radioactive materials and their decay products in the crust indicates that the crust as a whole can be considered to be 3 to 3.5 billion years old. Shmidt, using another method based on his own theory of the earth's origin, proposed an age of nearly 7 billion years for the earth. At the present time, 4 to 4.5 billion years is the figure usually accepted for the earth's age.

PART THREE

THE CHIEF STRUCTURAL FORMS
OF ROCKS

CHAPTER 5

PRIMARY STRUCTURAL FORMS

It has been said that rocks form various geometric bodies in the earth's crust that are called *structural forms*. These structural forms are the basic material from which the geologist deduces not only the present structure of this or that part of the earth's crust but also the tectonic history of the region and the movements and changes that have taken place there in the past.

This chapter will briefly review the main types of structural forms and some of the complex structures that have developed from them. More detailed information on the morphology of structural forms can be found in texts on structural geology (B. Willis and R. Willis, 1932; Leith, 1935; Billings, 1946; Nevin, 1942; de Sitter, 1956).

Structural forms may be classified as *primary* and *secondary*. Primary structures are those which originated at the time the rock itself was formed and are closely associated with the conditions of formation. Secondary structures appear as a result of mechanical deformations of the rocks after they were formed; that is, they are due mainly to tectonic movements.

This simple principle of classifying structures is not always easy to apply. For example, if a layer of sedimentary rock is deposited during a subsidence of the earth's crust within a fairly limited area, the beds themselves are downwarped as they are deposited, finally forming a structural basin in accordance with the movements of the earth's crust. For each bed, taken separately, such downwarping is secondary, since each bed initially had an almost horizontal position. For the series of beds as a whole, however, this structural form may be regarded as primary, since it originated at the time the beds were formed. To avoid confusion, all the structural forms considered as primary and all those considered as secondary will simply be enumerated below.

Primary structural forms depend on the conditions under which the rock was formed and thus are generally different in sedimentary and igneous rocks. Metamorphic rocks, since they are predominantly altered sedimentary rocks, generally have structural forms characteristic of the latter.

PRIMARY STRUCTURAL FORMS OF SEDIMENTARY ROCKS

The basic structural form of sedimentary rocks is *bedding*. One exception is the irregularly shaped masses formed by reef limestones within strata of other sedimentary rocks; their number, however, is not great enough to disrupt the generally regular structure of the sediments that contain them. Many publications have been devoted to describing and classifying the bedding of sedimentary rocks. These problems have been examined in detail by N. B. Vassoyevich (1951, etc.), who defined bedding as "all primary changes in the lithology of sediments in the vertical direction (in section and in time), regardless of any lateral changes within the stratigraphic unit (in the horizontal direction), as long as they are flat and relatively thin."

Thus bedding is expressed as an alternation of various rocks in a section. Adjacent beds are often similar in composition; for example, bedding may be observed within a generally homogeneous stratum¹ of clay, limestone, etc. In such cases, even when the adjacent beds are similar, some difference in composition or structure may be observed between the top of the underlying and the base of the overlying bed. Clastic material, for example, is often coarser at the bottom of a bed than at the top; limestones are purer at the bottom of a bed than at the top, where a small admixture of argillaceous material can be found in them, and so forth. The bedding is clearest where the rock changes markedly from one bed to another, as in an alternation of beds of limestone, sandstone, clay, etc.

Bedding is determined not only by greater or smaller differences in the composition and texture of the rock, but also by the presence of dividing surfaces between the beds. These surfaces do not always take the form of visible cracks, but they can generally be seen when the rock is deformed by natural causes or split artificially. Beds may vary considerably in thickness, from several millimeters to several meters.

Sometimes, in studying geologic sections, one can see an uninterrupted sequence of beds, which can often be verified by paleontologic data. In other cases there are breaks in the sequence of deposition. These are the so-called stratigraphic breaks, which take the form of a hiatus in some stratigraphic subdivision. Such breaks are usually indicated morphologically by traces of ancient erosion. These traces destroy the top surface of the bed lying directly beneath, cutting it up into uneven hollows and bulges that are usually small but sometimes have an amplitude of a few score meters. Rocks deposited after the break

¹The word "stratum" will here be used for a group of beds of unspecified size. The term "suite" will refer primarily to strata that have a definite stratigraphic position. "Layer" (*plast*) will mean a bed or small number of beds whose lithology does not change along the strike.

usually contain products of the erosion of underlying rocks (basal conglomerates).

The sequence of rocks associated with bedding is seldom disorderly. Ordinarily it is possible to determine both the sequence of the beds in the section and the distribution of beds of greater or lesser thickness. In one place, for example, thick strata of thin-bedded clays can be seen, and in another part of the same section or in another section there is a predominance of thick sandstone beds, while a third section contains mostly massive limestones or an alternation of limestone and clay, and so forth. *Cyclic bedding*—a regular repetition of a limited “assembly” of rocks—is also known. This repetition is characteristic of flysch suites, in which each cycle is made up of sandstone or coarsely clastic limestone at the bottom, marl in the middle, and clays at the top. Such cycles can also be seen in many coal-bearing strata, where there is a regular alternation of sandstone, limestones, clays, and coal. Figure 8 shows an example of such cyclic bedding.

The regularity in the distribution of the rocks in the bedded sedimentary mantle of the earth’s crust makes it possible to group them by strata that differ from each other in the character and the sequence of their “assembly” of rocks. These strata are called *formations*. Some examples of formations are the flysch, or the stratum of massive Upper Jurassic limestones in the Northern Caucasus, or the suite of red sandstones and clays with interbedded gypsum and dolomite in the Permian sediments on the western slope of the Urals. It will be seen below that the distribution of sedimentary rock formations in the crust of the earth is governed by certain laws that are of great practical importance.

Another essential feature of sedimentary rock strata is their change in the horizontal direction. By tracing any stratum of rock that has a definite place in the stratigraphic section, one sees that the stratum undergoes changes in various places: these changes encompass both the lithologic composition and the thickness. The changes in lithology will be called *changes in facies*, or *facies changes*, “facies” here being taken to mean the lithologic composition of the rock.

Change in the facies and thickness of sedimentary strata is a general rule; thus the composition and thickness of a given stratigraphic subdivision (stage, series, system, etc.) never have more than local significance. The changes occur at various rates. Sometimes the facies and thickness are constant over a large area, and changes can be observed



FIG. 8. An example of cyclic bedding (after N. B. Vassoyevich): the Sabuinskiy flysch along the Khandokhevi River in the Caucasus.

only after hundreds of kilometers. In other cases the changes are very sharp: the general appearance of a stratum and its thickness and composition will change abruptly within a few kilometers and sometimes within even hundreds of meters. The Upper Jurassic clays on the Russian plain, for instance, are noted for the uniformity of their facies over an enormous area; only in the Donbas, the Crimea, or the northern Caucasus is the Upper Jurassic represented by other rocks (limestones). The thickness of these sediments in the central regions also changes slowly and very little; it remains on the order of several scores of meters. But in the Caucasus the Upper Jurassic is hundreds, and in places more than a thousand, meters thick.

Figure 9 shows an example of rapidly changing facies and thicknesses.

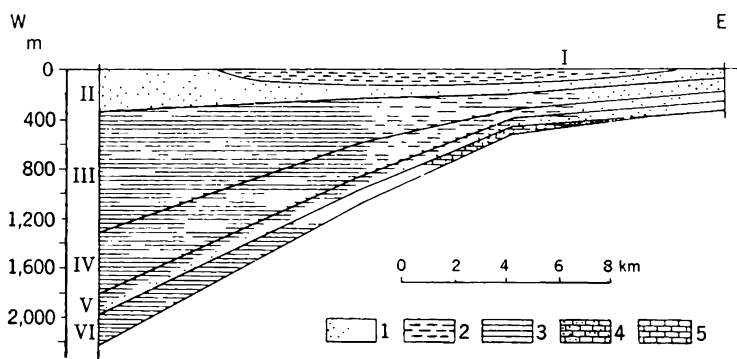


FIG. 9. Change in the thickness and facies of the Lower Cretaceous sediments in the northern Caucasus: (I) Albian; (II) Upper Aptian; (III) Lower Aptian; (IV) Barremian; (V) Houterivian; (VI) Valanginian; (1) sands; (2) sands and clays; (3) clays; (4) sandy limestones; (5) limestones (after V. V. Beloussov).

On the Belaya River in the northwestern Caucasus, the Lower Cretaceous is made up almost exclusively of coarse sands about 300 m thick. Eight kilometers to the west the Lower Cretaceous is represented by alternating sandstones and clays to a thickness of 500 m. Six kilometers farther to the west the same suite has turned entirely to dark clays and siderites and reaches a thickness of 1,000 m. Finally, another 10 km to the west, at the Pshekha River, the thickness of the Lower Cretaceous clays increases to 2,200 m. All these changes can be traced step by step, so that there is no possibility of the existence of any tectonic faults.

The changes in the facies and thicknesses of sedimentary strata are of great importance. Failure to take account of these changes has sometimes led to serious errors in interpreting the tectonic structure of the Alps. Seeing different facies and thicknesses close to each other in rocks of the same age, geologists concluded that enormous overthrusts and great horizontal displacements of portions of the earth's crust had

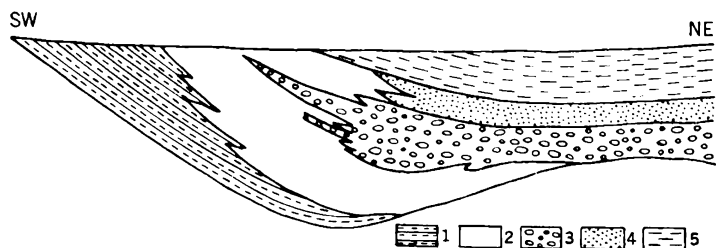


FIG. 10. Diagram showing the relationship of facies in the Ciscarpathian frontal flexure: (1) sand and clay of the Polyanitsa suite; (2) salt-bearing suite; (3) conglomerates of the Sloboda suite; (4) sandstones of the Dobrotov suite; (5) sandstones, clays, and marls of the Stebnikov suite (after M. V. Muratov).

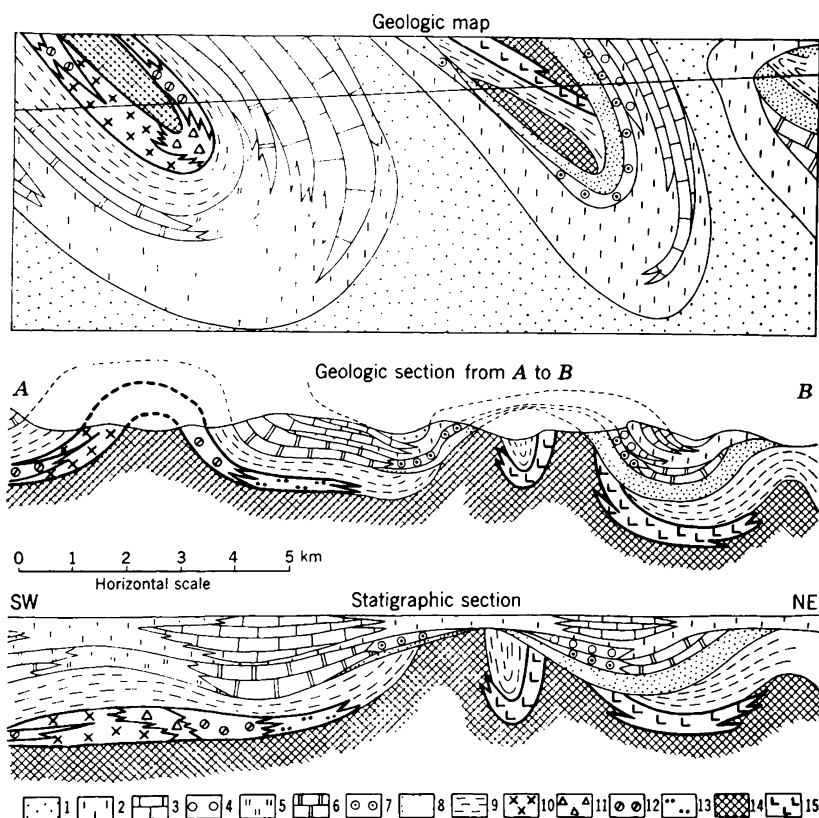


FIG. 11. Greatly simplified diagram of the changes in facies and thickness of the Middle Paleozoic rocks in the Baydzhansay anticlinorium of the Kara-Tau ridge (after M. V. Gzovskiy): (1) Upper Tournaisian (sandstones and clay shales); (2-4) Lower Tournaisian; (2) massive limestones, (3) bedded limestones, (4) limestone conglomerates; (5-7) Famennian; (5) massive limestones and dolomites, (6) finely bedded limestones and dolomites, (7) carbonate conglomerates; (8-13) upper part of the Silurian-Devonian; (8) red sandstones and conglomerates, (9) clay shales, (10) acidic lavas, (11) volcanic breccias made up of acidic lavas, (12) volcanogenic conglomerates of acidic lavas, (13) arkosic sandstones; (14-15) lower part of the Silurian-Devonian; (14) graywackes and clay shales, (15) porphyritic extrusives.

brought together rocks that originated far from each other, in completely different sedimentary environments and tectonic zones. In reality, no such huge tectonic movements ever took place in the Alps, where there were only sharp but quite normal changes in the facies and thicknesses of the rocks in certain areas (Belousov, Gzovski, and Goryachev, 1951).

Both sudden and gradual changes in facies parallel to the bedding surface usually occur as an interfingering of the facies in a transitional zone. Figure 10 shows a diagram of a normal facies change. Separate beds of one group of rocks wedge into the beds of the other facies. The zone of facies change is usually irregular both in the horizontal view and in the vertical section. An inclined position of the zone of transition in the vertical section indicates movement of the zone in time. Figure 11 shows a geologic map and geologic sections of an area in which the changes in the facies and thicknesses of certain series were determined and mapped.

Sometimes a given facies is very limited in extent, occupying a narrow strip between other facies. If this facies is also limited in time, it

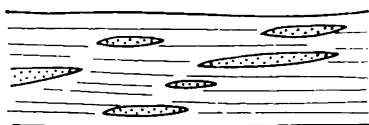


FIG. 12. Sand lenses in clay.

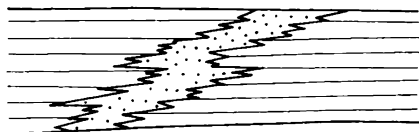


FIG. 13. Diagram showing the replacement of clay by a narrow zone of sand.

will not form an extensive flat bed in the section but rather short lenses wedging out rapidly in two or more directions. Figure 12 shows a section with isolated lenses of sand within a stratum of clay and Fig. 13 a section in which a sandy facies cut across clay beds in a narrow strip but was maintained for a considerable length of time, while a thick stratum of sediments was accumulated. The width of this strip of sand varied in time and changed its location in space.

PRIMARY STRUCTURAL FORMS OF IGNEOUS ROCKS

It is known that igneous rocks include both volcanic, or extrusive, and plutonic, or intrusive, rocks, whose primary structures differ according to their different modes of origin.

The primary structural forms of *extrusive rocks* are in most cases similar to those of sedimentary rocks. Lavas pouring out on the earth's surface form *flows*, which resemble beds. Layering can also be seen in the loose products of volcanic eruptions: volcanic ash and bombs. Volcanic ash forms layers of tuff, and bombs form layers of volcanic breccia.

Loose volcanic products are often washed down by rain soon after their eruption and mixed with clastic sedimentary material, thus forming mixed volcanic-sedimentary rocks (tuffites) that have a bedded structure much like that of the more common sedimentary rocks. Because of their layered structure, extrusive rocks are commonly interbedded with sedimentary rocks. One exception is the extremely viscous lavas of acidic composition, which do not flow far from the center of eruption but are concentrated around it in the form of a volcanic dome; such domes, however, are rarely found in geologic sections.

Like sediments, extrusive rocks make up certain formations in the earth's crust that differ from each other in composition, in eruptive origin, and in other characteristics. These formations also differ in the relationship of volcanic and sedimentary rocks. Areas abounding in extrusive rocks contain some formations composed primarily of basic flows and *ejectamenta* (porphyritic formations) and others of acidic lavas (porphyries). There are also many formations of combined tuffaceous and sedimentary origin and of lavas interbedded with sedimentary rocks, etc. In geologic literature one frequently encounters the term "ophiolitic formations"; these are characteristic of certain areas and consist of sand and clay strata with many interlayers of basic lavas.

Like sedimentary strata again, extrusive rocks also change laterally in composition and thickness. Lava flows frequently change laterally into tuffs and breccias. When the products of simultaneous eruption from different centers are adjacent in a section, they may all have different compositions, forming a zone in which lava flows of different compositions will alternate and wedge into each other.

At the very center of a volcanic eruption one may sometimes see an earlier volcanic vent filled with solidified lava or breccia in the form of a distorted cylinder. Such cylindrical extrusive bodies are called *necks*. As one moves away from the center of eruption, the extrusive rocks wedge out.

The most common primary structural forms of *intrusive rocks* are batholiths, dikes, sills, so-called minor intrusives, and scattered injections.

Batholiths are large igneous bodies of acidic composition, whose shapes are not accurately known because only their uppermost parts can be seen. Until quite recently it was assumed that batholiths are roughly conical, pyramidal, or columnar in shape; it was also assumed that these bodies increase in width as they extend downward all the way to the bottom of the earth's crust and that they are directly connected to regions below the crust from which the magma emanates—hence the frequently used name—*bottomless (abyssal) intrusions*. Recently, however, these views have been revised. Although there is no question of the immense size of batholiths, many writers suggest that

they are not bottomless and are contained wholly within the upper parts of the earth's crust. Batholiths are known for the uniformity of their chemical composition. They are always made up of acidic rocks: granites, granodiorites, and rarely diorites. The relatively homogeneous composition of the rocks in a single batholith is also characteristic. In Proterozoic or younger strata the batholith usually cuts across the intruded structures; in Archean strata, however, batholiths usually have a conformable contact with the enclosing rocks.

Fissure intrusions, or *dikes*, were formed by the penetration of magma into the gaps that appeared as tectonic movements cracked the earth's crust. A fissure intrusion forms a tabular body, which is either vertical or at a steep angle within the earth's crust. The thickness will vary

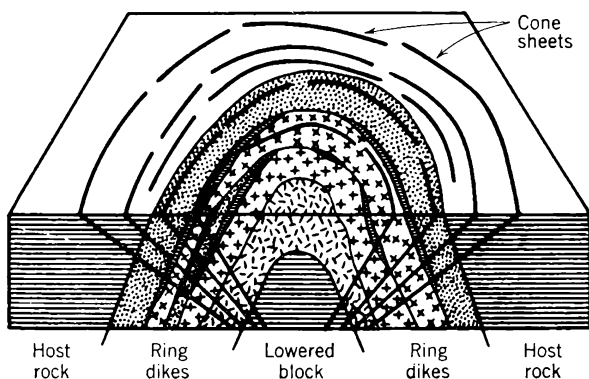


FIG. 14. Diagram of ring dikes in Scotland. The horizontal lines indicate intruded rocks (after Richey).

from a few centimeters to a few score meters, and in rare cases to hundreds of meters. Such bodies can be traced along their strike sometimes for scores and even hundreds of kilometers, though not always in a straight line. Discontinuous and curving intrusions can be found, and in Scotland and the United States one may see peculiar systems of ring dikes and cone sheets with the apex at the top or at the bottom, respectively. Figure 14 shows a diagram of such ring dikes and cone sheets (Richey, 1948). Dikes vary greatly in composition, from the most acidic to the most basic rocks.

Sills are tabular igneous bodies that have been intruded parallel to the bedding planes. Some sills cover thousands of square kilometers and maintain an astonishingly uniform thickness over their entire area. In New Jersey (U.S.A.) there are Triassic sills 300 m thick that may be traced for 160 km (Daly, 1933). Sometimes a number of sills interbedded with sedimentary rocks may be seen within a single section. Sills occasionally have irregular shapes resulting from an uneven lifting

of the overlying rocks as they were intruded. Their composition can vary, but as a rule they are made up of basic rocks.

"Minor intrusives" is an inclusive term covering a variety of small intrusive bodies. This includes so-called stocks, irregularly cylindrical bodies that intersect the structure. This class also encompasses lens-shaped and mushroom-shaped bodies that have been intruded between beds and lifted them up to form domes: these are *laccoliths* (Fig. 15).

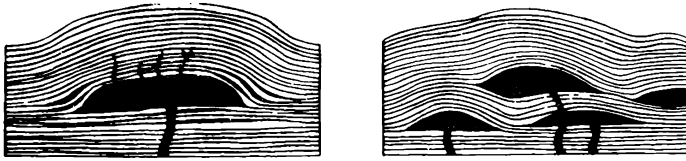


FIG. 15. Laccoliths of various shapes.

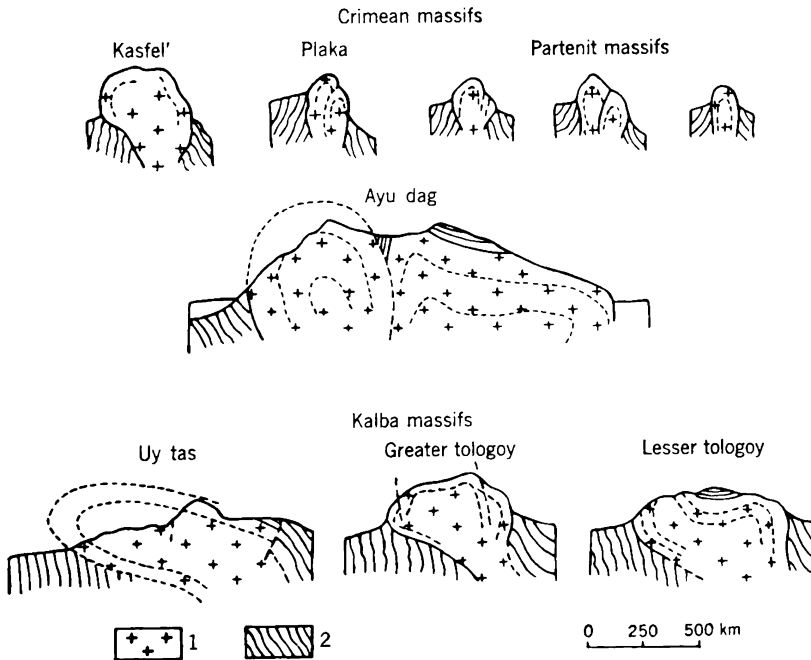


FIG. 16. Drop-shaped intrusives: (1) intrusive rock; (2) host rock (after V. N. Pavlinov).

Igneous bodies with the shape of recumbent drops also belong to this group. V. N. Pavlinov has studied and described intrusive bodies of this shape in various regions of the U.S.S.R. (Pavlinov, 1947); he has determined that many laccoliths are not mushroom-shaped, as was believed earlier, but more closely resemble drops. This is true of laccoliths in the Crimea, the Caucasus, the Kalba, and the Urals (Fig. 16). These

intrusives are, indeed, quite small in comparison to batholiths: their outcrops are no more than a few kilometers in diameter, whereas batholiths are sometimes a few hundred kilometers across. Stocks cut across the structure of the enclosing rocks, whereas laccoliths make conformable contacts with them. Drop-shaped bodies have conformable contacts at the top, but below they cut across the structure. Minor intrusives vary greatly in composition and include many alkaline rocks: trachytes, syenites, monzonites, shonkinites, etc.

Scattered injections are widespread in Precambrian strata but much less common in younger rocks. They take the form of systems of very small (even microscopic) *lit-par-lit* injections, which often saturate the invaded sediments to such an extent that it is difficult to distinguish the igneous from the sedimentary rock. The composition of scattered injections is acidic, and their total volume in Archean strata is enormous.

CHAPTER 6

SECONDARY STRUCTURAL FORMS: COHESIVE DISLOCATIONS

Secondary structures are the result of mechanical action by external forces on rocks. It has been agreed that mechanical processes of deep-seated origin that change the structures of rocks will be called *tectonic movements*. There are some processes, however, that change the structures of rocks but cannot be regarded as tectonic movements; these include fracturing due to weathering, landslides, caving, the mechanical action of glaciers, tunneling out of limestones by ground waters, erosion by water, changes in the volume of rocks by hydration, and so forth.

A change in the structure of a rock that results from tectonic causes is called a *dislocation*; thus the secondary structural forms of rocks are called *dislocated structures*.

Under the action of tectonic forces the geometric bodies formed by the rocks in the crust of the earth change their shapes and form new geometric bodies. When such changes take place, the rocks may maintain their cohesiveness without fracturing or they may be fractured; that is, fissures and faults may be formed in them. In the first instance the dislocation is plastic, or *cohesive*; in the second case it is *disruptive*.

The final shape of the structural form, of course, if the deforming forces remain the same, will depend on the primary structure and the mechanical properties of the deformed rocks. The secondary structures produced in bedded sedimentary rocks, extrusive rocks, and sills are different from the structures that arise under identical conditions in massive intrusive rocks. This difference is due, in the first case, to the nonhomogeneous bedded structure of the deformed strata and, in the second case, to the massive structure and irregular external shape of the igneous rocks.

It must be remembered that it is not as easy to detect secondary structures in massive rocks as it is in bedded rocks. The presence of bedding makes it possible to see the details of the deformation in all parts of the dislocated rocks, whereas it is much more difficult to recognize the dislocation in massive homogeneous intrusives.

The secondary structures that can be seen in layered rocks—bedded sedimentary rocks, extrusive rocks, metamorphic rocks, and sills—will be examined first; manifestations of tectonic disturbances in massive intrusive rocks will be discussed briefly later. Plastic dislocations will be considered in this chapter, in which the following elementary types of cohesive dislocation are distinguished: distortion, bending, large downwarps and upwarps, and folds.

DISTORTION

Distortion, or change in the thickness of beds under the influence of tectonic forces, is a frequently encountered phenomenon. Such changes

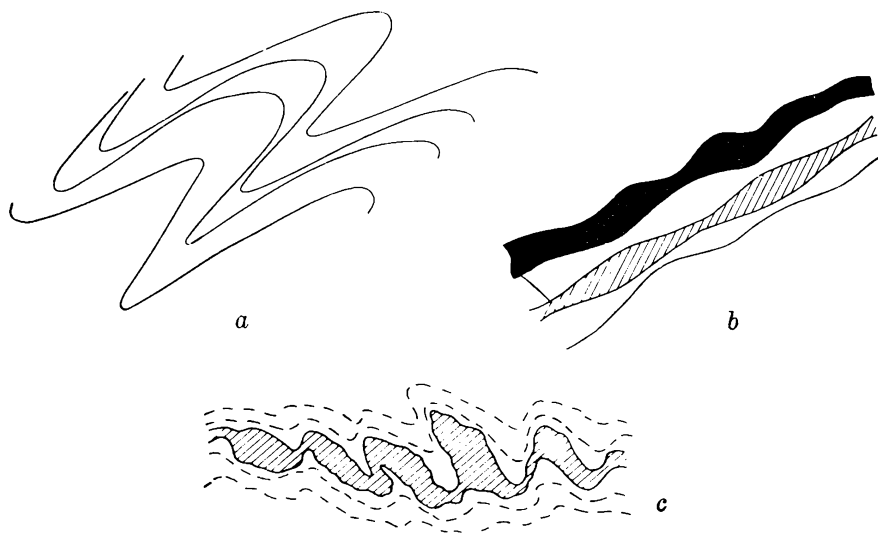


FIG. 17. Examples of distortion of the initial thickness of beds: (a) changes in the thickness of beds that have been crumpled into folds (decrease in the limbs and increase in the crests and troughs); (b) beds squeezed and pressed into lenses; (c) beds squeezed into lenses and crumpled into folds.

may occur in a bed over a great area or as local changes in the original thickness of the bed. In the first case, only a *decrease* in the original thickness can be observed; in the second, one may observe thinning of the bed, wearing through, squeezing out, necking, etc. Regular secondary *increases* in the thickness of a bed over a large area have not been observed.

In most cases local changes in the primary thickness of the bed involve alternate increases and decreases in the thickness. One may thus speak of a periodic sequence of alternate squeezing out and bulging of the bed. If such changes occur with periodic regularity, the bed may

be divided into tectonic lenses; in non-Soviet literature the term *boudinage* is used to designate this phenomenon. Figure 17 shows some examples of this type of distortion.

Much more pronounced and much larger distortions are found in nature. These are typical of the most plastic rocks (clay, salt, gypsum, and anhydrite) and appear under specific tectonic conditions to be considered below. Such distortions are produced by extreme flowage

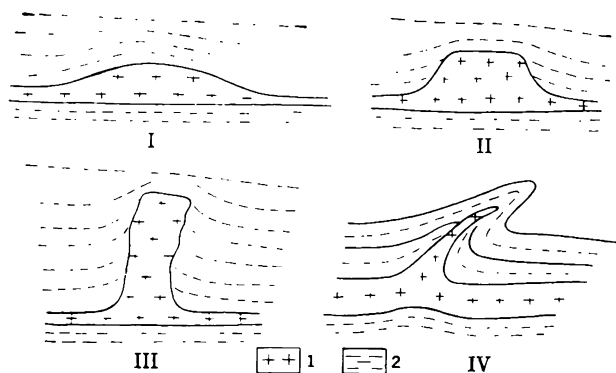


FIG. 18. Shapes of diapir bodies: (I) thickening of a stratum of salt; (II) stump-shaped diapir body; (III) columnar body; (IV) comb-shaped body; (1) plastic rocks (salt, clay, etc.); (2) surrounding rocks.

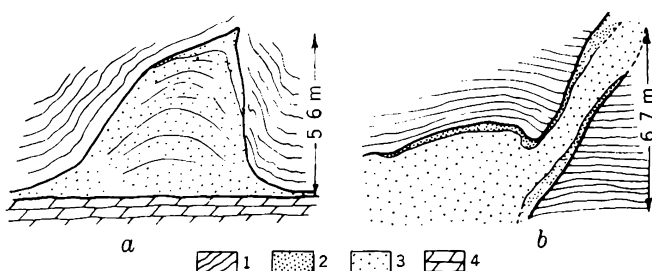


FIG. 19. Neptunic dikes. Intrusion of the Danian sands into the rocks of the Lower Syzranian series (Middle Volga region): (1) Lower Syzranian rocks; (2) quartz sandstones; (3) sands; (4) Upper Cretaceous marls (after V. V. Bronguleyev).

of plastic rocks. They include diapir cores, in which the rocks are subjected to extremely strong distortion, being squeezed out from some places and accumulating as swellings in others. Typical diapir cones resemble steep or gently sloping peaks, cones, cylinders, or columns (Fig. 18); but a great variety of shapes are found, running the gamut from gently sloping peaks and conical or dome-shaped surfaces to thin columns several kilometers high with a diameter no greater than 200 to 300 m. Swellings in diapir cores may be hundreds of meters, or even kilometers, in extent. Diapir columns may be vertical or inclined; the

upper part is often thicker or even mushroom-shaped. In some cases, they taper downward toward the base, forming an inverted teardrop. In cross section the diapir columns are round or elliptical.

The same class includes so-called *neptunic dikes*, formed as a result of the forcing of plastic rocks into relatively narrow fissures (Fig. 19). Neptunic dikes occur not only in regions of dislocation but also in areas in which the beds remain undisturbed (Bronguleyev, 1947).

BENDING

A bend is an abrupt change in the attitude of a bed. The lines that indicate the location of bends are *hinge lines*; these follow the bend within the limits of the bed (Fig. 20). If the hinge line is horizontal, the bend is vertical; if the hinge line is vertical, the bend is horizontal. Slanting bends whose hinge lines are inclined to the horizontal are

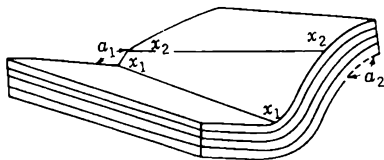


FIG. 20. Two bends in a group of beds forming a flexure: x_1 , x_2 , hinge lines; a_1 , a_2 , angles of bends.

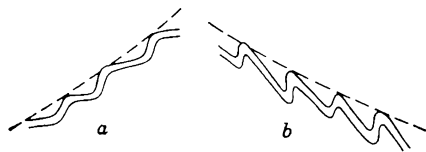


FIG. 21. Stairways of flexures: (a) flexures running along the level; (b) flexures running counter to the level. Dotted lines indicate the levels of the flexures.

most often encountered. The size of a bend, like that of any dislocation, is limited; if one traces a bend laterally in both directions, one will always arrive at points at which the dislocation ends.

A bend in a bed is always part of a more complex dislocation consisting of a combination of several bends. Combinations of two bends, which are known as *flexures* (Fig. 20), are often encountered in nature. A flexure is a steplike structure beyond which the bed returns to its former attitude.

A flexure is defined in space by the positions of its two hinge lines: if both hinge lines are horizontal, the flexure is vertical; if the two hinge lines are vertical, the flexure is horizontal. In the first case the flexure will appear in profile in the vertical section, and in the second the flexure will appear in full profile in the horizontal plane. Most often, however, flexures are inclined. The flexure may be complicated by the fact that its two hinge lines are not parallel; as a rule, they have different orientations in space.

Flexures are also characterized by the attitudes of their two outer limbs and their inner limb and by the size or height of their limbs. A

vertical flexure has a lower and an upper outer limb. The bends of a flexure are characterized by the angles between the inner and the two outer limbs; these angles may differ considerably. A flexure with right angles, though it is often shown in diagrams in textbooks, is very rare in the field. Flexures of very different sizes occur, varying from a few centimeters to two or three kilometers in their dimensions.

Flexures are often encountered in groups, which may form a series of steps of vertical and inclined flexures (Fig. 21). In this case, the imaginary surface connecting either the upper or the lower bends is called the *level of the flexures*. Flexures may run along the level or counter to it. In the former case, if one moves along the steps formed by the flexures, the inner limbs are inclined in the same direction as the level of the flexures: downward when the level slopes downward, and vice versa (see Fig. 21a). When the flexures run counter to the level, the inner limb is inclined in the opposite direction to that of the level: upward when the level moves downward, and vice versa (see Fig. 21b).

LARGE DOWNWARPS AND UPWARPS

By large downwarps and upwarps are meant those extremely great downward- and upward-bending structures of the beds forming the earth's crust that, because of their great size, cannot be called folds.

Two types of large-scale downwarps and upwarps are recognized: (1) *synclises* and *antecclises*, and (2) *synclinoria* and *anticlinoria*.

Syneclises and Antecclises

Syneclises and antecclises are gently sloping and extensive downwarped and upwarped structural forms occurring on platforms. They usually take the shape of irregular circles or ovals. To the extent that alternating synclises or antecclises cover the area of a platform, the opposite structures, which have less regular outlines, fill up the remaining space between them.

Syneclises and antecclises may be hundreds and even thousands of kilometers in diameter; the depth or height of the downward or upward curvature is relatively small (never exceeding 5 km, and more often only 2 or 3). Thus the dip of the beds is exceedingly small and may be determined only by a regional survey over a large area.

An essential peculiarity of synclises and antecclises is the fact that the dip of the beds on the flanks differs between the deeper beds and those close to the surface. In both synclises and antecclises, the angle of dip diminishes from bottom to top; the overlying beds dip more gently than the deeper ones. This is due to the regularly changing thickness of the deposits in the structure, which are thickest at the center of a synecclise and correspondingly thinnest at the center of an

antecline. The decreases in the thickness of certain strata in antecises often continue to the point where they are wedged out, so that the stratigraphic sequence of antecises is less complete than that of synclises. As a rule, there is also a facies change; the facies change to coarser, littoral types toward the centers of antecises. Some facies may also be replaced by others: limestones, for example, which are well developed at the center of a syncline, may give way to clays or sandstones as one moves toward the edges (see Fig. 22). All the changes in thickness and facies are governed by the process of formation of these structures.

The Moscow basin is a typical example of a syncline. Here the crystalline Precambrian basement lies at a depth of about 1,700 m below the surface; it is covered by Lower Paleozoic and Devonian deposits about 1,300 m thick, which in turn are overlain by Carboniferous rocks

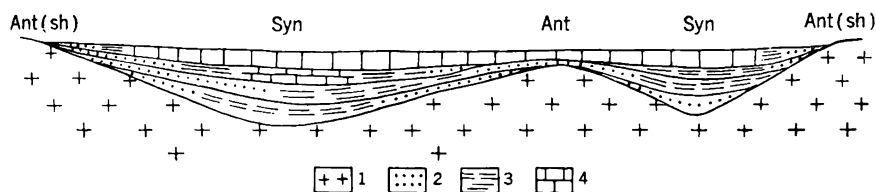


FIG. 22. Diagram of a typical section through synclises (*Syn*) and antecises (*Ant*), including shields (*Ant/sh*): (1) crystalline basement; (2) sandstone; (3) clay; (4) limestone.

about 300 m thick and by Upper Jurassic and Lower Cretaceous sediments about 100 m thick. Toward the northwest the crystalline basement begins to rise gradually; about 700 km from Moscow it crops out on the surface on the slopes of the Baltic shield. Thus this rise amounts to 2.5 m/km, forming an angle of less than 10 minutes of arc. The strata of sedimentary rocks also rise with the crystalline basement, in the same direction. In this direction the thickness of the beds decreases, individual series gradually wedge out, and at the shores of Lake Ladoga the entire complex of sedimentary rocks wedges out. Similar changes are observed to the south and west. Although here the sedimentary rock series does not disappear entirely, individual suites wedge out and the total thickness of the sediments is greatly reduced; in the vicinity of Orel, for example, the sedimentary rocks are less than 500 m thick. The thickness decreases to the same degree in the west, near Vitebsk. Along with the decrease in the thickness of the deposits toward the periphery of the syncline, there are also changes in facies. Thus the Devonian and Carboniferous marine sediments at the center of the basin to a considerable degree change to continental facies toward the edges of the syncline (especially to the north and west).

Another example of a syncline is the eastern Russian depression, where the Paleozoic and Mesozoic sedimentary deposits are more than 2,000 m thick in places. In the Pericaspian depression to the south they are even thicker.

The crystalline Baltic shield, the Ukraine massif, and the Voronezh massif are examples of anteklises. On the Baltic shield, which is an extensive convex bulge, the Precambrian crystalline rocks form outcrops on the surface, and sedimentary deposits are completely lacking. The thickness of the sedimentary rocks thus decreases toward the center of the anteklise to such an extent that they have been wedged out completely. The Ukraine massif has only a thin cover of Tertiary deposits over the uplifted crystalline basement; there are no Paleozoic and Mesozoic rocks, although they are deposited to a great thickness in the adjacent Donets basin. On the Voronezh massif the blanket of sediments is only 150 to 200 m thick, and the stratigraphic section is incomplete: Paleozoic rocks are totally absent, or represented only by Upper Devonian or Carboniferous deposits. Of the Mesozoic there are individual strata of Upper Jurassic and Upper Cretaceous rocks, and the Tertiary is represented only by part of the Paleogene.

Synclinal and Anticlinal

Synclinal and anticlinal are downwarps and upwarps of much greater amplitude and contrast than synclises and anteklises; their outlines are also different. These structures are associated with folded zones and appear as elongated ovals in maps or when seen from above. They may be several score kilometers but are usually hundreds of kilometers long and from a few kilometers to a few score kilometers wide. The amplitude of the downwarp or upwarp is always at least several kilometers and may reach ten or twelve kilometers. These structures are therefore much more pronounced than synclises and anteklises. A juxtaposition of synclinal and anticlinal presents much greater contrast than a juxtaposition of the previously described structures. A typical feature of synclinal and anticlinal is the presence of a number of secondary folds (ordinary anticlines and synclines) within their structure; synclinal and anticlinal are, in fact, a combination of these secondary folds with an overall downward or upward bending.

The main Caucasus range, the Ural-Tau range, and the Sikhote Alin' range are examples of anticlinal; all these are huge anticlinal structures. Examples of synclinal are the Kura depression, the Rion depression, the Fergana Valley, and the Tadzhik depression. The synclinal character of these structural forms is revealed in the fact that the youngest sediments (Tertiary, in the above examples) are contained in the center of the structure, whereas the margins are made up of older, Mesozoic rocks.

Anticlinoria and synclinoria climb and plunge along their strike. The most deeply plunging parts of anticlinoria are the ends; the ends of synclinoria, on the other hand, represent the culminations of their axes.

The anticlinorium of the main Caucasus range plunges into the earth's surface in the southeast at the Apsheron Peninsula and in the northwest at the Taman Peninsula. The surfaces of these plunges are covered chiefly with Tertiary deposits, which also extend across the axis of the anticlinorium. The culmination of this axis is situated on the line from the Mineral Waters area to the Dzirula crystalline massif. Near the culmination the anticlinorium widens considerably, and the oldest rocks crop out at its center.

The Kura and Rion synclinoria farther south plunge and rise in the same directions. The first of these plunges southeastward toward the Caspian Sea; it widens in this direction, and the youngest deposits—Upper Tertiary and Quaternary—predominate at the surface. The Rion synclinorium plunges westward toward the Black Sea. Both synclinoria rise in the direction of the Dzirula massif, where the most ancient crystalline rocks that form their basement emerge from under the Tertiary and Mesozoic deposits and crop out at the surface. The Kura and Rion synclinoria may actually be considered as a single synclinorium that culminates at the Dzirula massif and plunges away from it on both sides.

The structures of anticlinoria and synclinoria are often asymmetrical: one flank is steeper than the other. An example of this is the anticlinorium of the main Caucasus range, whose southern flank is much steeper and shorter than the northern, where the rocks dip gently northward.

To understand the structures of synclinoria and anticlinoria properly, one must remember that their shapes will differ in different horizontal sections. One must also distinguish between *noninverted* and *inverted* synclinoria and anticlinoria (Gzovskiy, 1948). In noninverted structures the angle of dip of the beds decreases from the bottom to the top; this has also been noted in the case of syneclises and anteklises. Here the deeper strata also dip more steeply than the higher beds (Fig. 23); in addition, as in the previously mentioned structures, some of the beds wedge out completely toward the center of an anticlinorium, whereas there is a more complete stratigraphic section in the synclinorium. The facies also tend to change, becoming coarser shore-line types as one approaches the center of the anticlinorium.

The structure of inverted synclinoria and anticlinoria is more complicated. Here the lower-lying strata may have a structure the reverse of that of the higher ones. Anticlinoria formed by the beds near the surface may thus be superimposed on synclinoria formed by the deeper strata, and vice versa. These discrepancies result from the peculiar dis-

tribution of thicknesses in the sedimentary strata that form inverted anticlinoria and synclinoria. In structures of this type the sedimentary strata may be divided into two series. In the upper series, the thicknesses change in the same way as in noninverted structural forms, increasing in the synclinoria and decreasing in the anticlinoria. In the lower series the strata become thicker under the anticlinoria and thinner under the synclinoria. Thus, in areas where inverted anticlinoria and synclinoria are situated next to each other, the sedimentary strata form large lenses within the crust of the earth. These lenses are on two levels, upper and lower, and are staggered with respect to each other: in the higher level the thicker parts of the lenses are underneath synclinoria, whereas in the lower level they are under anticlinoria (see Fig. 23). As

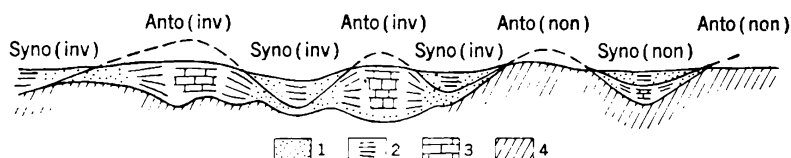


FIG. 23. Diagram of the structures of synclinoria (*Syno*) and anticlinoria (*Anto*), inverted (*inv*) and noninverted (*non*): (1) coarse clastic rocks; (2) clays; (3) limestones; (4) older complex of rocks.

a result, the attitudes and structures of the rocks in the different levels are determined by the relationship between the height of the anticlinorium or depth of the synclinorium and the differences in the thickness of the sediments beneath. If the downwarp (the increase in thickness) of the strata in the lower series under an inverted anticlinorium is great enough in relation to the height of the anticlinorium, the lower strata will form a downwarp, or synclinorium (see Fig. 23). If, however, the height of the anticlinorium is greater than the downwarp (or bulge) in these strata, they will form an anticlinorium that is smaller and more gently sloping than the one above. The possible variations in inverted synclinoria may be seen in Fig. 23.

The facies changes in inverted synclinoria and anticlinoria are also different in the upper and lower levels: in the lower series the coarser facies originating closer to the shore line are found nearer to the synclinoria. In the upper series the reverse is true: the coarser facies occur on the limbs of anticlinoria, as in the case of noninverted structures. In other words, in both cases the appearance of coarser facies parallels the decrease in the thickness of the deposits. All these changes, again, are closely connected with the conditions of formation and the history of noninverted and inverted synclinoria and anticlinoria.

A number of examples are given below. Noninverted and inverted

structures may be seen in the Caucasus (see Fig. 23): the main Caucasus range, taken as a whole, is an inverted anticlinorium. The Upper Tertiary deposits on its flanks dip away from the axis of the anticlinorium at all points; their thickness, in addition, decreases toward the axis. Only over the plunges of the anticlinorium—toward the southeast and the northwest—do these deposits completely cover the structure, crossing the axis from one limb to the other. In the greater, central or culminating, part of the anticlinorium they wedge out on the flanks and do not reach the crest. The facies of these sediments also change, becoming coarser in the upper parts of the flanks.

A somewhat different picture emerges from the reconstruction of the lower strata of Jurassic deposits within this anticlinorium. In the center of the granite Dzirula massif, south of the main Caucasus range, the bottom of the Jurassic deposits is greatly elevated hypsometrically; it does not rise toward the main range but descends considerably in this direction, because of the great increase in the thickness of the Mesozoic deposits below the anticlinorium. In the axial part of the range, the bottom of the Jurassic deposits again rises to the surface, but the general structure of the lower strata of Jurassic sediments beneath the main Caucasus range indicates that there is not only an anticlinorium but also a deep synclinorium, which is complicated at the center by a narrow secondary anticlinorium. In the eastern part of the Greater Caucasus, the bottom of the Jurassic deposits does not crop out at the surface at the axis of the range; here it is doubtless more downwarped than in the western Caucasus. The thickness of the Jurassic, Cretaceous, and Lower Tertiary (Eocene) sediments increases toward the center of the range. In all these deposits, coarser facies appear as one moves away from the axis of the range and approaches the troughs of the surrounding depressions.

The Kura depression is an inverted synclinorium, containing Upper Tertiary and some Paleogene deposits of great thickness. The greater part of the Mesozoic sediments, however, here decreases sharply in thickness; some of these deposits wedge out completely. Thus it may be inferred that the deeper strata under the Kura depression form a gentle anticlinorium.

A second inverted anticlinorium is located south of the Kura depression, in the Somkhēt mountains of the Lesser Caucasus. Here the thickness of the Jurassic deposits increases rapidly toward the axis of the anticlinorium, and for this reason the structure of the lower strata differs sharply from that of the higher beds.

There is a noninverted anticlinorium south of Lake Sevan in Armenia; this is the so-called Miskhan anticlinorium. All the deposits, from Jurassic to Tertiary, decrease in thickness and wedge out as one approaches the axis.

FOLDS

Folds are very common in stratified sedimentary rocks. Courses in general and structural geology present detailed morphologic classifications of folds based on the angle formed by their limbs, the plunge, the shape of the fold in the horizontal section, etc. On the assumption that the reader is familiar with this elementary classification, this section will describe another morphologic classification based not only on certain morphologic features of individual folds but also on complexes of folds occurring regularly within the earth's crust. This classification is important in understanding the conditions under which various types of folds have been formed.

Types of Folding

Continuous, or Holomorphic, and Discontinuous, or Idiomorphic, Folding. In studying the morphology of individual folds, especially as

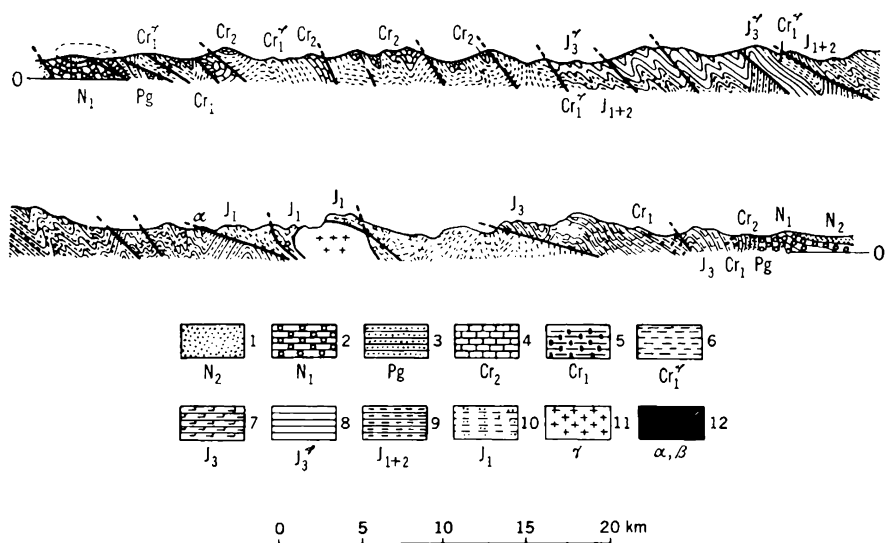


FIG. 24. Example of a continuous folded structure: (1) Pliocene; (2) Miocene; (3) Paleogene; (4) Upper Cretaceous; (5) Lower Cretaceous; (6) Lower Cretaceous (flysch facies); (7) Upper Jurassic; (8) Upper Jurassic (flysch facies); (9) Middle and Lower Jurassic undifferentiated; (10) Lower Jurassic; (11) granites; (12) andesites and basalts (north-south section along the Georgian route, central Caucasus) (after V. P. Rengarten).

seen in the field, it becomes readily apparent that the various folds may be divided morphologically into two basic types. These two types of folding have long been established but have been given a variety of names by different authors. The first type of folding has been called

Alpine, orogenic, geosynclinal, or linear; the second has been called Germanic, cratonogenic, platform, or domical. N. S. Shatskiy (1945) called the second type of folding *placanticlinal* and *placosynclinal*.

Folding of the first type is considered to be the result of deformation

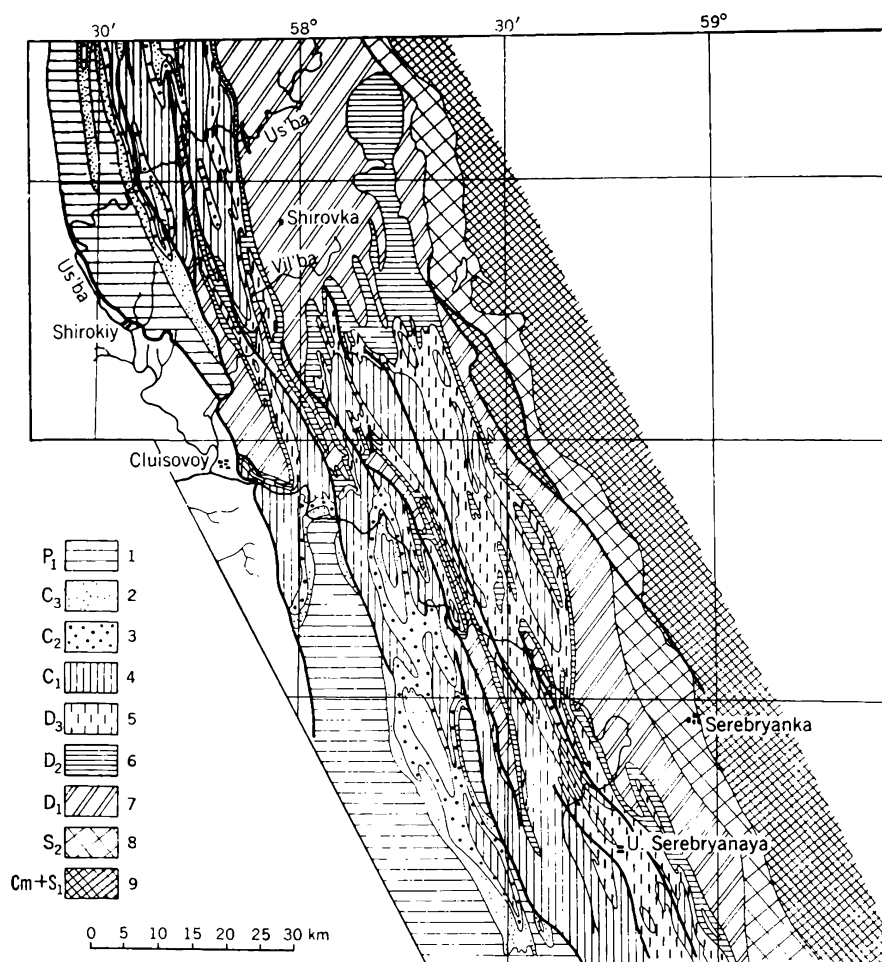


FIG. 25. Geologic map of part of the western slope of the northern Urals. An example of a continuous folded structure: (1) Lower Permian; (2) Upper Carboniferous; (3) Middle Carboniferous; (4) Lower Carboniferous; (5) Upper Devonian; (6) Middle Devonian; (7) Lower Devonian; (8) Upper Silurian; (9) Cambrian and Silurian.

of more plastic rock masses by horizontal forces of compression. These folds are long and tightly folded, parallel, and typical of folded zones. Folding of the second type is considered to originate in the dislocation of relatively "rigid" masses; thus, in addition to plastic bending,

there is a considerable amount of faulting in the formation of such folds, which are commonly less pronounced, have a greater radius and usually a more circular domical outline, and are less regularly disposed than the folds of the first type.

The terms proposed earlier do not convey all the special features of the basic types of folding. Thus it has become necessary to develop a new terminology that will more completely reflect the morphology of the folded structures and, if possible, the conditions under which they were formed.

Figures 24 and 25 are a structural section and a map of folded structures of the first type. These show a tightly folded structure that is continuous within the given zone. The folds follow each other in a

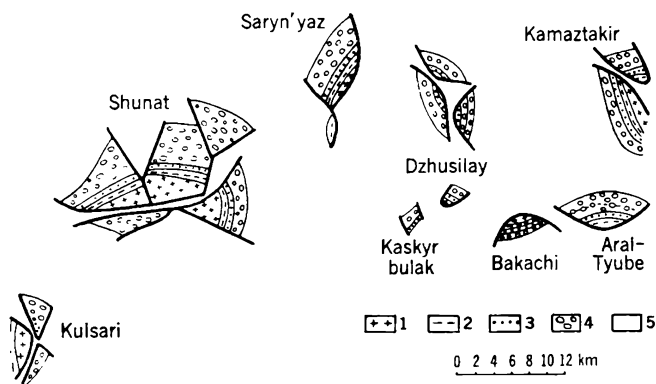


FIG. 26. Geologic sketch map of part of the Ural-Emba region. An example of discontinuous folding: (1) Jurassic; (2) Neocomian; (3) Aptian; (4) Albian; (5) Turonian-Cenomanian and younger deposits.

generally regular succession of anticlines and synclines. Here the individual fold is not a separate tectonic structure but forms part of the whole: the strike of each fold follows the general trend of the entire folded structure. The inclination of its axis and its shape in the horizontal and vertical sections are also determined by the overall structure. Thus it appears that this folded structure is created by a generally oriented horizontal movement of the rocks encompassing a large area.

A completely different picture is presented by regions in which the second type of folding appears: there are numerous isolated, unconnected, variously shaped uplifts in an area of horizontally lying beds. The Emba dome, the Oka-Tsna uplift, and the Black Hills of North America are examples of this type of folding.

Figure 26 shows a geologic map of part of the Pericaspian depression. The structure of this area is characterized by a number of strongly faulted domes, which have been called "broken plates." These domes

contain outcrops of rocks from the Lower Jurassic to the Lower Cretaceous; between the domes are horizontal beds of Upper Cretaceous and younger rocks. The isolated nature of these uplifts is shown clearly on the diagram. It may be noted in passing that the folding is represented here only by anticlinal structures. The areas between them are not true synclines but remnants of depressions consisting of horizontal beds.

The structure of a part of the Volga-Ural region will serve as a second example (Fig. 27). At first glance, Fig. 27 appears to show an area of

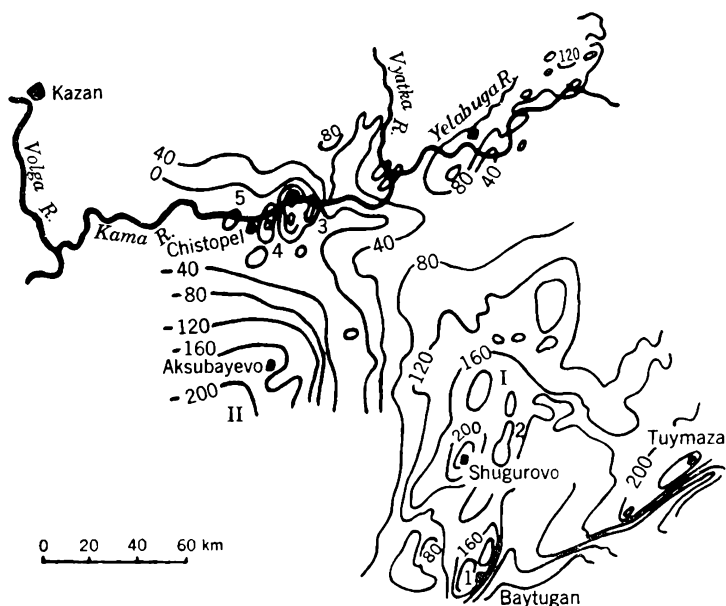


FIG. 27. Structural map of the southeastern part of the Tatar Republic and the adjacent parts of the Bashkir Republic and the Kuybyshev region: (I) Sok-Sheshma elongated uplift; (II) Aksubayevo trough; (1) Baytugan dome; (2) Romashkino dome; (3) Buldyr dome; (4) Zmiyev dome; (5) Kubass dome. Structure contours drawn along the top of the *Spirifer* horizon (after Mel'nikov).

linear folding with a predominantly northeasterly trend. A closer examination, however, will reveal that the general trend of the group of folds does not affect the character of each individual fold.

In the southern part of this area, near the tributaries of the Sok, the Sheshma, and Stepnoy Zay Rivers, there is a broad uplift known as the Sok-Sheshma arch, which is broken up into a number of folds (domes) of a secondary order. Is there any linear trend in the disposition of these domes? The general trend of the uplift is north-northeast; the individual domes have the same trend. Each dome, however, is a separate unit, and one cannot group the domes along any common line. In the south, for

example, is the easily distinguishable Baytugan dome; if one follows this along the trend to the northeast, it plunges away and is replaced in this direction by part of the Bugul'ma depression. If one similarly follows the trend of the Romashkino dome (see Fig. 27), one encounters a "structural nose" to the southwest; continuing in this direction, one comes to the center of a structural depression. North of this dome along its trend there is a small uplift, but still farther north the traces of this "line" disappear. In the Shugurovo area the Sok-Sheshlma uplift is obscured by two rows of secondary domes; there is a single dome along its trend to the northeast, but the structure contours show no connection with the two lines of domes above. Moving to the part of this region near the Kama River, one finds a number of domes: at the mouth of the Uzh River, near Naberezhnyye Chelny, at Yelabuga, and at the mouth of the Vyatka River. All these domes are arranged along a line extending toward the northeast; but if one follows their trend toward the southwest, one finds that they disappear in this direction.

The isolated nature of these domes is emphasized by the fact that their actual structures consist only of anticlines. Where one would expect synclinal folds, there are only residual depressions. These depressions, however, are not independent structures; they are determined entirely by the shape and location of the adjacent domes.

The following terms, based on an analysis of the examples given above, are suggested as designations for the two types of folding. Because of their common shape and, apparently, their common origin, as well as their disposition in close parallel lines, folds of the first type—Alpine, geosynclinal, or linear—may be called *continuous*, or *holomorphic*, folds.¹ Groups of such folds form a continuous, or holomorphic, folded structure. The isolated nature of folds of the second type—Germanic, platform, or domical—suggests the term *discontinuous*, or *idiomorphic*, folds.²

Characteristic Features of the Two Basic Morphologic Types of Folds

Continuous, or holomorphic, folding has the following characteristics:

1. *Continuous distribution* over the entire folded area: folds completely cover the area, and there are no portions without folding.
2. *Equal development of anticlines and synclines*: the equal width and complexity of the positive and negative structural forms are average qualities, but they are nevertheless a characteristic feature of this type of folding.

¹ The Greek word *holos* (ὅλος) means "whole," "total," "complete"; thus this term conveys the unity of the folded structure, the common shape and origin of the folds, and the manner in which they continuously cover an area.

² *Idios* (ἴδιος) in Greek means "partial," "individual," "isolated"; this term thus suggests the isolated, separate nature of these folds.

3. *Linearity*: all the folds have the same common trend. Changes in the trend appear not in single folds alone, but in whole groups of folds.

4. *Horizontal, oriented movement* of the rock masses, manifested in the same regular inclination of the axial planes over a large area. Within a considerable territory these axial planes are inclined or overturned in the same direction.

Typical discontinuous, or idiomorphic, folding has the following characteristics:

1. *Discontinuous or local nature* of the folds: single and isolated folds situated in an area of horizontal beds. The folds resemble isolated uplifts, domes, brachyanticlines, or flexures.

2. *Unequal development of anticlines and synclines*: in typical cases the actual structures are represented only by anticlines (or domes), and there is a total absence of corresponding synclines; instead there are residual depressions with horizontally lying beds. If the anticlines are close to each other, a synclinal structure appears to be formed between them. These synclines disappear, however, if they are traced for any distance, and closer observation shows them to be subordinate structures completely determined by the shape and location of the anticlines. The synclines grade into residual depressions in the directions of their trends.

3. *Absence of linearity*, manifested either in the different trends of individual folds or, even when a number of folds have parallel trends, in the absence of any direct connection between their trends.

4. *Absence of horizontally oriented movement* of the masses of rock; that is, the axial planes of different folds are inclined in different directions.

It should not be concluded, on the basis of the above characteristics, that there is no regularity in the distribution of the folds in discontinuous folding. Such regularities exist; they will be discussed below along with the origin of this type of folding. They are more complicated, however, than the simple linearity that is the most prominent feature of holomorphic folding.

Morphologic Peculiarities of Discontinuous Folds

Discontinuous folds have a much greater variety of shapes than continuous folds, which look alike and differ only in size, in the degree of compression, in the angle of inclination of their axial planes, and in the nature of their disharmony. Thus a number of morphologic varieties of discontinuous folds will be described briefly, without any attempt at classification.

Many discontinuous folds have the shape of *domes*. These are domical upwarps of the beds and are circular or oval in outline. The beds forming the flanks dip at angles from 5 to 30°. The domes may vary in

diameter from a few score meters to a few score kilometers; the amplitude of the uplifted beds may also vary considerably. The beds of some domes remain unbroken and unaffected by faults whereas in other cases the domes are broken by normal and reverse faults that have lowered or elevated parts of the dome. Figure 28 shows an example of a dome that has been only slightly fractured by faults, and Fig. 29 shows one that is strongly faulted.

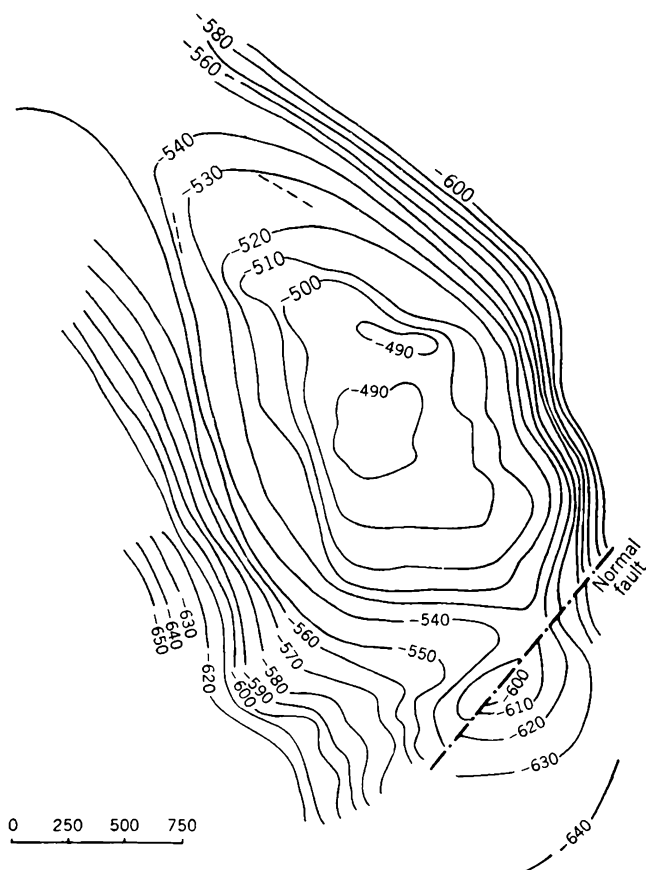


FIG. 28. Structural map of a simple dome (United States).

Among discontinuous folds there are a number of box folds, both unilateral and bilateral. Such structures as the Zhiguli and Don-Medveditsa dislocations belong to this type, as do many of the uplifts in the Rocky Mountains of the U.S.A. These are large massive upwarps with the shape of brachyanticlines. Their lengths are measured in scores of kilometers; the amplitude of these uplifts is hundreds of meters, and sometimes more than a kilometer. The domical shape of

the upwarp is broad and smooth but usually complicated by gently sloping secondary domes. The flat tops are contrasted with the steep flanks, which thus resemble flexures. Sometimes one flank forms a sharply defined flexure, while the other slopes gently downward (in unilateral box folds). An example of this is the structure of the Zhiguli uplift (Fig. 30a). In other cases both flanks are steep, as in the Don-Medveditsa dislocations (bilateral box folds, Fig. 30b). The flanks of such folds are typically complicated by stairs of flexures; the steeper flanks are also cut by reverse faults. Figure 31 shows sections through certain uplifts in the Rocky Mountains.

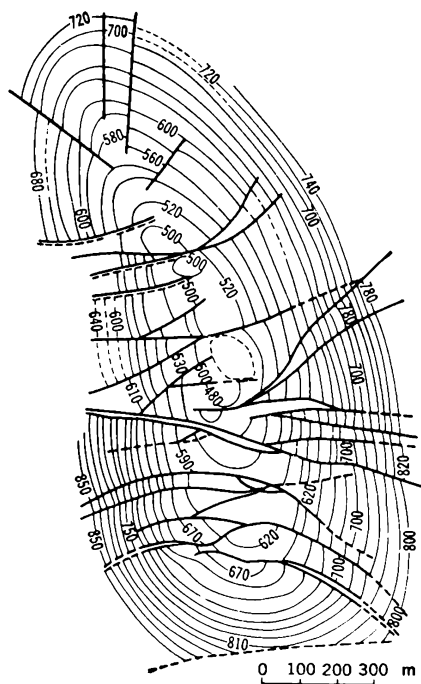


FIG. 29. Structural map of a faulted dome (Apsheron peninsula). The heavy solid lines are normal faults.

Another variety of discontinuous folding is the *elongated low uplift* whose length considerably exceeds its width. The flanks dip very gently, no more than a few degrees or even fractions of a degree. One example is the Oka-Tsna elongated uplift, which extends from the Klyaz'ma River in the north to the Tsna River in the south; its length is about 350 km, its width 25 to 40 km, and its amplitude no more than 200 to 300 m. The angle of dip of the beds on the flanks is no more than $1\frac{1}{2}^{\circ}$. Rocks of the Middle and part of the Lower Carboniferous crop out on its axis. On the eastern flank these are replaced by Upper Carboniferous and Permian rocks; on the western flank the Paleozoic rocks are covered by the Jurassic deposits of the Moscow basin.

Elongated uplifts often have extremely complex shapes in the horizontal section: they can be ameoboid, with offshoots, "noses," etc., or be obscured by numerous second- and third-order domes. There are many such elongated uplifts between the Volga River and the ridge of the Ural Mountains, such as the Tuymaza, the Sok-Sheshma, the Vyatka, and the Krasnokamsk uplifts and many others.

All varieties of discontinuous folding are characterized by the regularity with which the thickness and facies change. The thicknesses of the deposits that form the structure of the fold generally decrease toward

the top and increase on the flanks toward the depressions between the folds. Some beds wedge out entirely on the crest of the fold; in addition, the beds that form discontinuous folds generally flatten out toward the top of the stratigraphic sequence (Fig. 32). The facies, in turn,

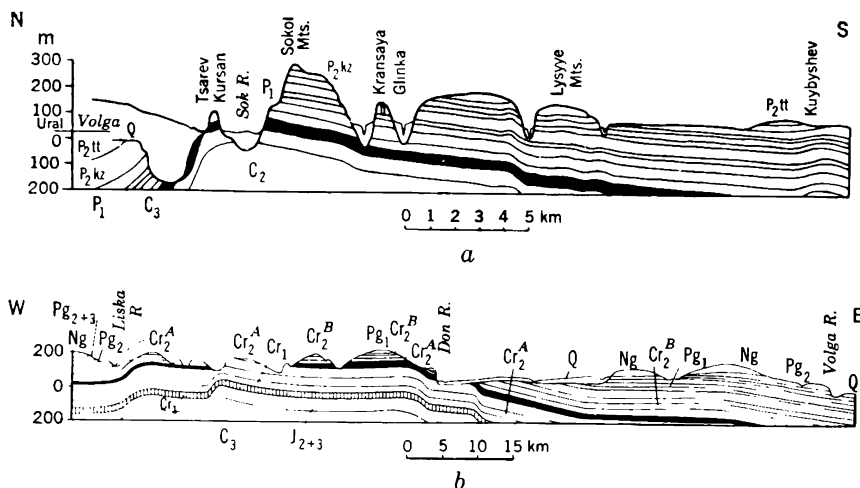


FIG. 30. Structural sections through the Zhiguli (*a*) and the Don-Medveditsa (*b*) uplifts. In section *a* the normal symbols are used. In section *b*, Cr_2^A = Cenomanian, Turonian, Coniacian, and Santonian; Cr_2^B = Campanian and Maestrichtian; Pg_1 = Syzranian and Proleyan beds; Pg_2 = Tsaritsyn beds; Pg_3 = Maykopian beds (after Ye. V. Milanovskiy).

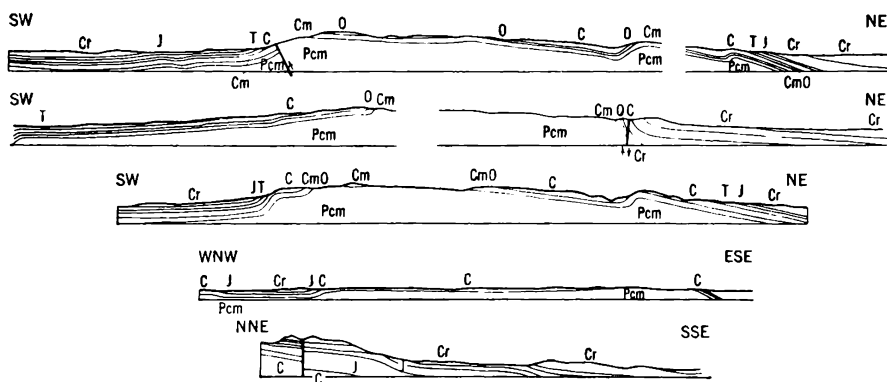


FIG. 31. Sections through the Rocky Mountains (after N. Darton).

tend to become coarser toward the crest of the folds. Thus discontinuous folds, in regard to their changing thicknesses and facies, resemble anticlines and anticlinaloria.

The domes in Kabristan in the southeastern Caucasus (Veber, 1935) are an example: according to V. V. Veber, there is a great difference

between the deposits in the domes and in the areas between the domes. In the depressions between the domes are argillaceous rocks forming an almost complete sequence of Neogene deposits, whereas near the crests there are much thinner sections of sandy rocks with many gaps in the stratigraphic column. The Don-Medveditsa uplift is another example. Here the thickness of all the deposits, beginning with the Middle Jurassic and ending with the Paleogene, decreases rapidly as one approaches the crest of the elongated uplift, and there is a transition to coarser facies. Some beds wedge out completely. The thicknesses and facies in many of the gently sloping elongated uplifts formed in the Paleozoic deposits of the eastern part of the Russian platform between the Volga and the Urals also show great changes. L. N. Rozanov (1949, 1957) described many examples. Many instances of change not only in thickness but also in facies have been noted in the domes and elongated uplifts of the Volga-Ural region: coarser facies become more extensive in the clastic sediments as one moves toward the crests of the uplifts. Such changes in facies and thickness, however, are not

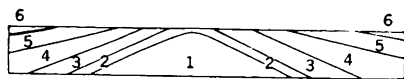


FIG. 32. Section through the Neftyanaya Gora dome (after K. P. Kalitskiy): (1) Red beds; (2 and 3) Lower Apsheron; (4) Middle Apsheron; (5) Lower Baku beds; (6) Upper Baku beds.

always observed in all the strata involved in the dislocations. Often these will be seen in the higher beds but not in those below a certain stratigraphic level. In other cases, strata in which there are no changes in thickness and facies alternate with strata in which such changes are observed.

This regular decrease in thickness toward the crest of a discontinuous fold in all or some of the strata generally causes the beds to dip more steeply in the deeper parts of the structure. Observations show, however, that there may be more complex differences in the attitudes of the various beds in discontinuous folds. In numerous cases the top of the dome has been displaced. Many cases are also known in which, in an area of discontinuous folding, the structure of the deeper strata has no resemblance to the structure of the higher strata, and the locations of structural uplifts and depressions may differ at different stratigraphic levels of the section. L. N. Rozanov's paper (1949) contains a number of examples of such structural discrepancies between various horizons. The Shugurovo dome at the top of the Kazanian stage (in the Volga-Ural region) has only one apex; along the top of the Ufa deposits, a slight bifurcation of the structure can be detected; the attitudes of the Carboniferous deposits, finally, clearly show the presence of two domes. In another instance, the Buldyr uplift in the Upper Permian deposits on the left bank of the Kama River has a simple structure. In the Artinskian beds the structure begins to bifurcate, and in the Carbonif-

erous deposits two peaks are clearly visible. Such increasing complexity in the structure of the fold with increasing depth, taking the form of a bifurcation of the fold in the deeper strata, can be seen in many discontinuous folds of the Transvolga area near Kuybyshev.

In addition to the increase in the steepness of the flanks and in the complexity of the dislocation, the opposite—that is, leveling out of the beds and simplification of the structure—has also been observed in many cases. In the Shugurovo dome, for example, although the structure becomes more complicated as one descends from the Kazanian stage to the Lower Carboniferous, the Devonian beds dip so slightly as to lie almost horizontal. The structure of the Yablonov dome (on the Malaya Kinel' River) first becomes more complex with increasing depth and then becomes simpler again: the Upper Permian deposits form one apex, the Lower Permian and Carboniferous form two, and the Devonian beds form only one again.

In certain cases the changes in the structure with increasing depth are more elaborate. The Aksubayevo dome (in the Tatar Republic) is symmetrical at the top of the Lower Kazanian deposits; in the Artinskian deposits its northern flank becomes much steeper. In the Tuymazin elongated uplift the crest moves toward the northeast and simultaneously toward the increasingly steeper flank, as the depth becomes greater. In many domes and elongated uplifts of the Transvolga area the crest of the dome is displaced, with increasing depth, both along the trend and perpendicular to it. In all these cases, the discrepancy in the attitudes of the beds at various stratigraphic levels is associated with irregularity in the distribution of the thicknesses of the individual suites. If the structure slopes more gently with increasing depth, for example, this means that the thicknesses of the beds increase toward the crest of the fold instead of becoming thinner in the usual manner (Fig. 33, lower). In other cases, structural discrepancies between the various strata are caused by more complicated changes in the thicknesses of individual suites within the fold (Fig. 33, upper).

It must be pointed out that pronounced discrepancies in the attitudes of the beds at various horizons have been observed only in very gentle discontinuous folds. Folds of this type that have a great amplitude and steep slopes everywhere show a regular decrease in the thicknesses of deposits going toward the crest of the fold.

There are transitional structures, as noted above, between anteklises and syneklises, on the one hand, and anticlinoria and synclinoria, on the other. There are also transitional forms between discontinuous anticlines and anticlinoria. In the latter case, the two structural forms are closely related, and it is hard to say when a structure ceases to be a large discontinuous uplift and begins to be an anticlinorium. An example of such transitional forms is the Mangyshlak, a domical upwarp composed

of Mesozoic and Cenozoic deposits, which is very large in comparison to ordinary discontinuous folds but quite small when compared with typical anticlinoria. The Great Balkhan, the Tuarkyr, and others might also be mentioned. Such transitional structures, which can be termed neither discontinuous folds nor anticlinoria, may be called *anticlinal uplifts*; the analogous transitional downwarps may be called *tectonic depressions*. It will be seen below that the transition between discontinuous folds and anticlinoria is determined by the similar origins of

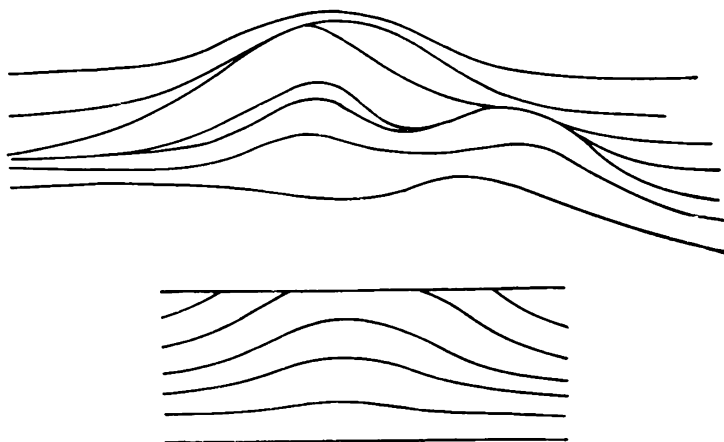


FIG. 33. *Lower*, section through a discontinuous fold sloping more gently toward the bottom. *Upper*, section through a discontinuous fold with discrepancies in the attitudes of the different beds.

these structural forms, which are so different from each other in their extreme examples.

Intermediate Folding (Ridgelike and Boxlike)

In addition to the two basic types of folding, many areas contain folded structures of intermediate character, which display features of both continuous and discontinuous folding to a greater or less degree. There are two varieties of intermediate or transitional folds; these will be called *ridgelike* and *boxlike* (or *box folds*).

Typical ridgelike folding is a succession of sharp, compressed anticlines (symmetrical, asymmetrical, or fan folds), often faulted by gently inclined or steep upthrusts, alternating with broad and gentle synclines. The folding is characterized, therefore, by a lack of similarity between the anticlines and synclines, but the synclines are not altogether absent, as in the case of discontinuous folding. Seen from above, this structure looks like a brachyanticline. Linearity and orientation of movement of the rocks are less pronounced than in continuous folding, but they can be noted to some degree.

An example of ridgelike folding is the structure of the Tera-Sunzha area in the region just north of the Caucasus. The section in Fig. 34 (Brod, 1940) presents a typical picture of broad gentle synclines and narrow faulted anticlines. There appears to be no orientation in the displacements along the faults, since in the last two anticlines this displacement is in opposite directions. It is well known that these anticlines (the Sunzha and the Tera) are brachyantoclines, when seen in horizontal section, and that they differ in shape from both typical linear folds and typical domes. A second example is the Jura Mountains on the border between France and Switzerland. The striking difference between the anticlines and synclines and the lack of any general orientation in the movement of the rock masses (that has been retained) may be seen easily in the section drawn by A. Heim (Fig. 35).

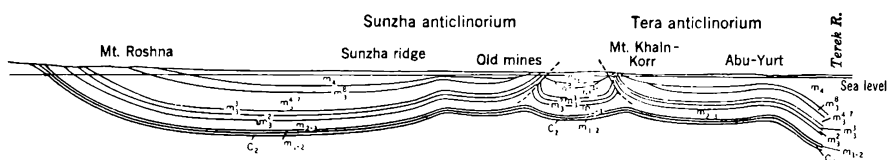


FIG. 34. Example of ridgelike folding. Section through the depression north of the Caucasus: m_4 , Postpliocene and Pliocene; m_2^s , Meotis; m_3^{4-1} , Sarmat; m_3^3 , Karagan beds; m_1^2 , Chokrak; m_2^2 , Maykopian; m_{1-2} , Foraminifera beds; C_2 , Upper Cretaceous (after I. O. Brod).

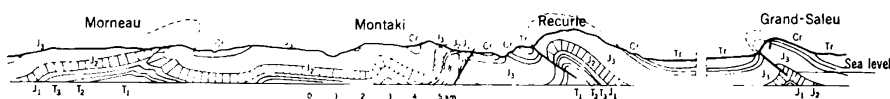


FIG. 35. Example of ridgelike folding. Section through the Jura Mountains (after A. Heim).

Boxlike folding looks like a mirror image or ridgelike folding; it is characterized by broad, massive, boxshaped anticlines with flat tops and steep, often vertical, flanks. The shape of the synclines depends entirely on the shape and location of the anticlines. If the latter are close to each other, the synclines between them will be compressed and narrow, resembling rifts. If the anticlines are at a great distance from each other, the synclinal depressions will accordingly be flat-bottomed and boxlike in profile. The limbs of the anticlines in boxlike folding are typically complicated by flexures, which sometimes form entire "stairways."

Boxlike anticlines have many morphologic features in common with discontinuous box folds. The difference is that the latter are isolated upwarps, whereas boxlike folding typically has a larger number of massive box folds formed over a large area. Boxlike anticlines are oval in horizontal section.

The folds in the Somkhet zone of northern Armenia are a good ex-

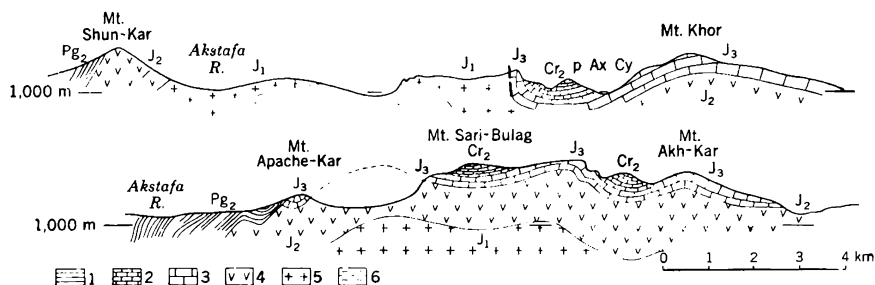


FIG. 36. Boxlike folds in the Akstafa River basin (Armenia): (1) Eocene; (2) Upper Cretaceous; (3) Upper Jurassic; (4) Middle Jurassic; (5) Lower Jurassic—quartz porphyries; (6) Lower Jurassic—tuffs (after M. V. Gzovskiy).

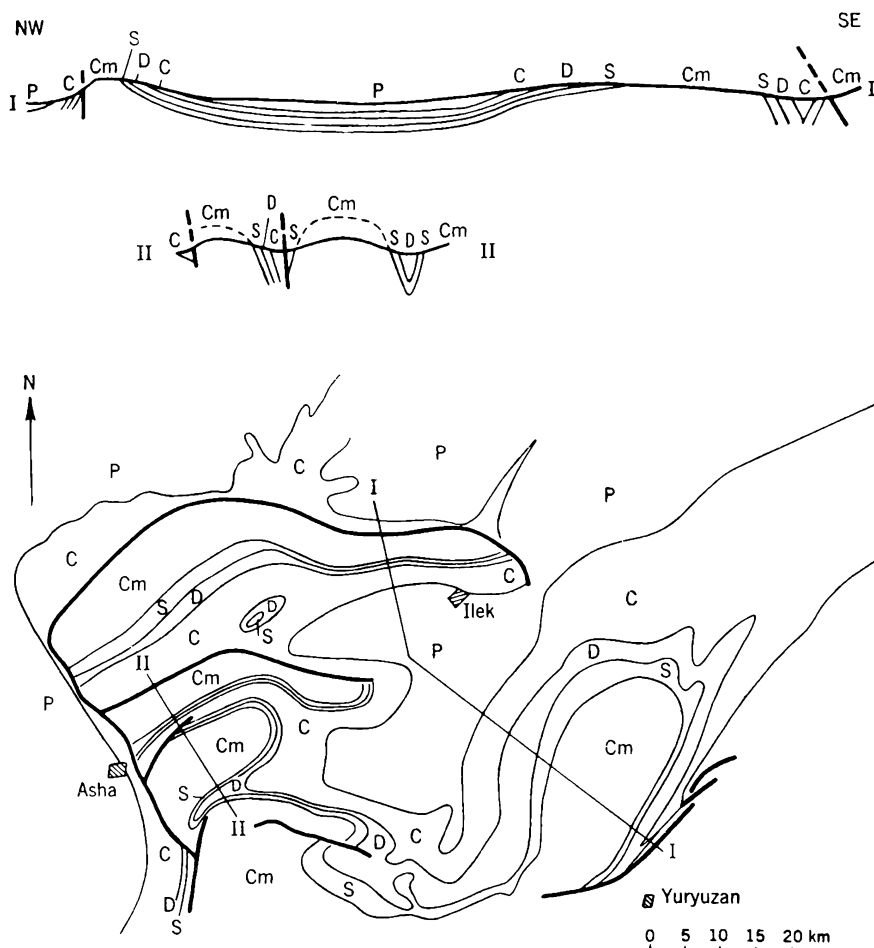


FIG. 37. Geologic sketch map and section through the southern margin of the Ufa plateau (Kara-Tau). Combination of ridgeline folding (I-I) and boxlike folding (II-II). Vertical scale exaggerated in the sections.

ample of boxlike folding. The outlines of the structure of this region are shown in Fig. 36. Similar features may be seen at the southern margin of the Ufa plateau (the Ural Kara-Tau), where boxlike folding is closely associated with ridgelike folding (Fig. 37). It will be seen that folding of the intermediate type is usually located between regions in which continuous and discontinuous folding predominate.

PLASTIC DISLOCATIONS IN MASSIVE INTRUSIVE BODIES

It was said earlier that extrusive rocks interbedded with sedimentary rocks, as well as intrusive sills, are deformed by tectonic movements essentially in the same manner as layered sedimentary rocks. Peculiarities that may arise in such dislocations are due only to the different mechanical properties of the various rocks. The behavior of massive bodies of intrusive rocks when subjected to tectonic deformations is more complex. Some researchers believe that massive intrusives, especially large ones, are not capable of plastic deformation and, when subjected to tectonic forces, will only fracture. This opinion is completely wrong. Observations in the field and in the laboratory have shown that, under appropriate conditions, intrusive rocks are sometimes capable of the same degree of plastic deformation without rupture as are sedimentary rocks.

It will be further seen that any solid body possesses plastic properties but that these plastic properties are manifested only under particular conditions of temperature, confining pressure, and, especially, rate of deformation. Because massive intrusive rocks are, in general, less plastic than layered sedimentary rocks, it sometimes happens that the same conditions that produce intensive plastic deformation in sedimentary rocks will cause massive intrusives to fracture. Under different conditions, however, at higher temperatures, at greater depths where the confining pressure is greater, and at slower rates of deformation, a massive intrusive may "flow" and be deformed to the same degree as the surrounding layered sedimentary rocks.

One must keep in mind, on the other hand, that the results of the deformation always depend on the initial structure of the deformed geologic body and on its internal composition. If a stratum of heterogeneous layered sedimentary rock is crumpled into folds by tectonic compression, it would naturally be futile to expect the same tectonic forces to produce the same structures in massive intrusive bodies that have an irregular original shape and a uniform internal structure. The results of the deformation, furthermore, which are easily seen in the folded structure of the sedimentary rocks, are much harder to detect in massive rocks, whose initial structure before the deformation is very difficult, if not impossible, to determine.

Let us assume that a massive intrusive body is surrounded on all sides by bedded sedimentary rocks (Fig. 38*a*), and that this portion of the earth's crust is then subjected to horizontal tectonic compression. As a result, the sedimentary rocks are crumpled into folds, as determined by their bedded structure (see Chap. 21). But the homogeneous structure of the massive intrusive is not capable of being crumpled and folded; thus it is deformed as a unit, being compressed horizontally and correspondingly elongated vertically. It is also clear that the intrusive is separated mechanically from the surrounding sedimentary rocks—there is failure along the contact between them. Such failure along the contact between massive and bedded rocks has actually been observed in the field in areas of dislocation and has sometimes incorrectly been interpreted as a large tectonic fault.

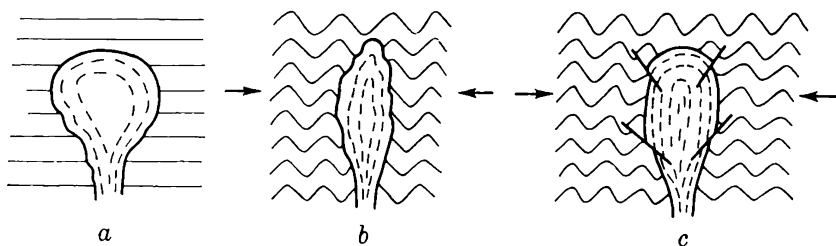


FIG. 38. Deformation of a massive intrusive surrounded by layered sedimentary rocks: (*a*) intrusive surrounded by undisturbed deposits; (*b*) the same intrusive subjected to compression simultaneous with the formation of folds in the layered country rock; (*c*) intrusive fractured under the same conditions.

The new shape of the massive intrusive body will differ from the original one, but this difference is extraordinarily difficult to establish reliably. Some help may be found in a study of the internal structure of the massive rock, which might reveal traces of internal tectonic dislocations in the form of a relocation and reorientation of the minerals. This will be discussed later. Figure 38*c* shows a massive intrusive body that, under the same conditions that deformed the surrounding rocks plastically, has turned out to be incapable of plastic deformation and has thus ruptured. This dislocation, of course, is easier to detect and analyze.

All that has been said here in regard to the deformation of massive intrusives is equally true of the deformation of the crystalline rock forming the "crystalline basement." There is a widespread opinion that if, for example, the sedimentary rocks on the Russian platform show numerous traces of plastic dislocation, the dislocations in the crystalline basement under the platform should take the form of faults and fractures. All this depends on the conditions that governed the development of the deformation: under certain conditions, or combinations of condi-

tions, the crystalline basement complex might prove to be more plastic than the sedimentary mantle and deform plastically, while the sedimentary deposits are fractured. Under other conditions the plastic dislocations in the sedimentary rocks will be accompanied by faults in the crystalline rocks.

Thus there is a variety of interrelationships between the deformations of sedimentary and ancient crystalline rocks in folded zones. G. D. Azhgirey (1951) has described a case of intensive plastic deformations in the crystalline basement of the northern Caucasus conforming to deformations in the overlying sedimentary strata: here the internal structure of the crystalline rocks is strongly affected by the deformation.

CHAPTER 7

SECONDARY STRUCTURAL FORMS: DISRUPTIVE DISLOCATIONS

Tectonic ruptures are divided into two basic groups: ruptures with displacement along the plane of rupture, and those without displacement. Ruptures without such displacement are called *tectonic joints*, or *diaclasses*; tectonic ruptures accompanied by such displacement are called *faults*, or *paraclases*.

M. A. Usov and his students (1933), after studying the structure of the Kuznets basin, use the term *disjunctive* for faults. The term *disjunctive dislocation*, however, was introduced at that time to designate all disruptive deformations, including both joints and faults; thus Usov's usage of the term is incorrect.

General texts on geology usually divide faults into strike-slip faults, normal faults, reverse faults, and overthrusts. This classification is inadequate, since it does not cover all the cases observed in nature. It is based on the assumption that the movement in a fault is always either along the dip or along the strike of the fracture. The strike-slip fault is considered to be a displacement along the strike and the other types of faults to be movements along the dip. Displacement in such simple directions, however, is actually seldom encountered: in most cases the displacement is in an oblique direction, being partly along the dip and partly along the strike. This combined character of the movement must definitely be taken into account in making any classification, since it determines the method of searching for the displaced strata. The main practical purpose of classifying faults is to serve as a guide in searching for the beds displaced by the fault. For this purpose the classification must be fairly complete. Such concepts, moreover, as normal fault, reverse fault, and overthrust are far from accurately defined, and the words are given different meanings by different writers.

Every tectonic rupture is, first of all, a fracture; in the case of a fault it is also a displacement along a *fault plane*. Fractures, or fault planes, are characterized by their position in space, their shape, and their size.

The rocks that border a fracture form its *walls*; if the plane of the fracture is inclined, there is a *hanging wall* and a *foot wall*. Fractures whose walls are in close contact are *closed*, whereas those whose walls are separated are *open*. The fissure formed by a fracture may remain empty for a certain distance, or it may be filled with mineral matter deposited by surface or ground waters or even with minerals of magmatic origin. Fissures may also be filled with the breccia resulting from tectonic fracturing or with the weathering products of the rocks forming the walls.

The main characteristics of faults are the direction of displacement and the amplitude. The displacement may be in any direction, but it will primarily follow the plane of the fault. As regards amplitude, several amplitudes must be distinguished for each fault.

Figure 39 shows a diagram of a fault along an inclined fracture. The displacement has followed neither the dip nor the strike of the fault plane, but has been in an oblique direction. This is the most common occurrence in faulting; in the terminology used in this book (see below) it should be called a *combined normal and strike-slip fault*. As a result of the dislocation, points *a* and *b*, which were together before, have moved apart. The distance *ab* is the total amplitude of the displacement, or the *net slip* (Billings, 1946). The line of the slip lies principally in the fault plane, and it may be determined by the striations, scratches, or grooves in its surface. Its direction may be expressed, as is usually done, in terms of the azimuth of the strike and the angle of dip, or in terms of the *angle of slip* *bac*; that is, the angle formed by the line of the slip and the strike of the fault plane.

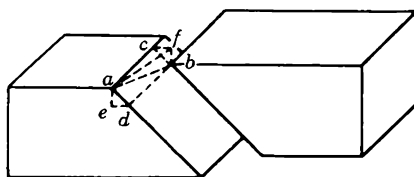


FIG. 39. Diagram of a combined normal and strike-slip fault.

The horizontal component *af* of the slip is the *horizontal amplitude* or the *horizontal displacement*. The distance *ac-db* is the strike slip and the distance *cb-ad* is the dip slip. The vertical line *ae-fb* should be called the *vertical amplitude* or the *vertical displacement* (the *throw*). Finally, for certain purposes it is sometimes necessary to determine the horizontal displacement in the direction perpendicular to the strike of the fault plane, or the line *ed-cf*: this is the *heave* of the combined normal and strike-slip fault.

Figure 40 shows a fault of another type, which belongs, in the terminology used herein (see below), to the category of *combined reverse and strike-slip faults*. The line *ab* is the net slip, the angle *cab* is the *angle of slip*, *af* is the *horizontal displacement*, *ce-bf* is the *vertical*

displacement, $ac-db$ is the *strike slip*, $ad-cb$ is the *dip slip*, and eb is the *heave*.

Since, as a result of the faulting (if the line of the slip does not lie in the bedding plane), beds of different age are in immediate contact with each other, one may speak of the *stratigraphic amplitude* of the displacement. This denotes the stratigraphic interval that is missing or is repeated in the section as a result of the displacement. This interval will depend, obviously, not only on the direction of the displacement but also on the attitude of the beds and the thicknesses of the individual stratigraphic subdivisions.

The same fault can have different directions and amplitudes of displacement at different places. As a rule, the amplitude is greatest at the middle of the fault line and decreases toward the ends, but there can also be local fluctuations in the direction and amplitude of the displacement.

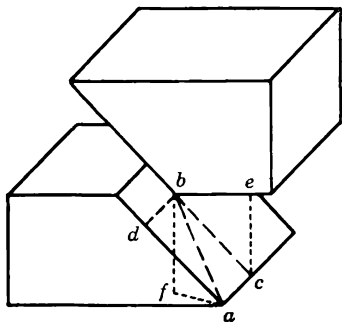


FIG. 40. Diagram of a strike-slip upthrust (explanation in the text).

From a practical point of view, the important result of faulting is the *repetition* or *omission* of beds. Repetition is seen when, after a bed is traced to the fault and a perpendicular to the bed is erected at that point, the same bed is encountered again along the direction of the perpendicular line on the other side of the fault plane (see Fig. 45). Omission is seen when the same bed is not encountered again after this operation (see Fig. 46). In the case of simple displacements,

the repetition and omission produced by a strike-slip fault will appear in the horizontal section, whereas those produced by the other types of faults will appear in the vertical section. In estimating the extent of stratified mineral deposits, the occurrence of repetition is a profitable factor, since it increases the mineral reserves in a given area. The omission of beds, on the other hand, results in decreased reserves, since part of the bed is missing within the boundaries of the district. Repetition or omission may be observed in the case of any fault, depending on the relationships between the direction of the displacement, the attitude of the beds, and the direction of the plane on which the observations are made.

JOINTS

Joints are extremely widespread within the crust of the earth. Observations have shown that there are almost no rocks in nature that are not broken by joints. The only exceptions are the most friable and

plastic rocks, in which the fractures are easily refilled as they form. In certain rocks, such as limestones, many of the fractures "heal"—they are rapidly filled by soluble material from the wall rock, which is re-deposited in them by ground waters. In different rocks, however, and in regions with different structures, the joints may vary greatly in size, shape, dip, position, and frequency.

The combination of fractures breaking up a given portion of the earth's crust is called the *joint system* of that area. The distribution of the joints determines the surface shape of a rock outcrop: when the jointing is distinct, the surface of the rock outcrop will usually be angular, with ingoing and outcoming solid angles, because the rock disintegrates primarily along the joints. Individual joints may differ in degree of separation, size, shape, and position in relation to the attitude of the beds, the trend of the folding, or the shape of a massive intruded rock.

According to the degree of their separation, joints are distinguished as latent, closed, and open. *Latent joints* are usually not visible, since they are extremely thin and hairlike, but they appear as the rock is broken up. They are an important factor, both favorable and unfavorable, in opencut mining, in driving mine tunnels, and in quarrying building stones. The presence of latent joints makes it easy to extract blocks of stone from quarries and, when the joints are favorably distributed, to obtain blocks of the desired shapes and sizes. On the other hand, latent jointing too frequently prevents the extraction of very large single blocks and decreases the strength and stability of mine tunnels. The danger of such jointing is that it sometimes cannot be detected in a preliminary inspection. *Closed joints* are those which are visible to the naked eye but have no noticeable separation. The walls of such joints are in direct contact, but there is no doubt of the fracture's existence. *Open joints*, as the name indicates, are characterized by a certain degree of separation. It should be noted that open joints are not always produced by motion of the walls away from each other. Often they originate from causes other than tectonic, for example, the widening of an earlier closed joint by weathering, which penetrates into the rock along the weakened surfaces. Capillary water can penetrate even strongly compressed fractures; this water, reacting with the rock, dissolves it and thus promotes its physical disintegration. This is the most common way in which closed joints become open.

It has been pointed out above that open joints can be sealed up again by filling with magmatic material, by deposits from aqueous solutions, or by the products of attrition of the rock.

In regard to their size, joints may be small and intraformational, or large and intersecting. *Small joints* are those which do not extend beyond the limits of a single layer; they are very widespread in the

earth's crust. In bedded rocks they intersect the individual beds, mainly perpendicular to the bedding. Thus none of the joints extends beyond the limits of the bed: the joint ends at the top and bottom of the bed, and in the next bed the joint running in the same direction is somewhat offset, located en echelon in relation to the first joint. Such joints also do not extend far in the direction of their strike, since they are limited in size and contained in all directions. *Large, or intersecting, joints* cut through a number of beds and in some cases are huge fractures extending for as much as several kilometers along both the strike and the dip. Joints have been known to extend for scores of kilometers along the surface of the earth. Such large fractures, however, are usually accompanied by displacement along the plane of fracture, thus becoming faults.

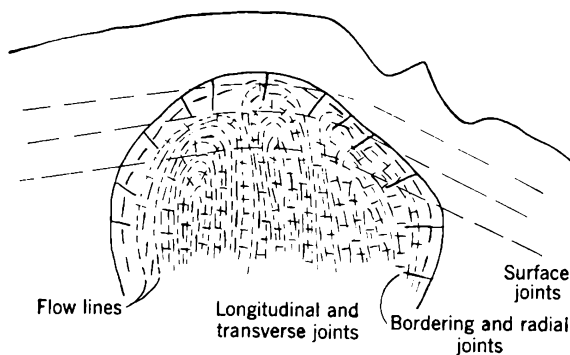


FIG. 41. Structure of a massive intrusive.

In relation to the attitude of the beds, joints may be perpendicular to the bed, parallel to the bed—that is, running within the bed along planes generally parallel to it—or at an angle to the bed. Perpendicular joints and joints forming an oblique angle, in relation to the bed's attitude, may be parallel to the dip, parallel to the strike, or diagonal. Diagonal joints are those whose strike is different from both the dip and the strike of the bed.

Relative to the trend of linear folds, joints may be *longitudinal*, *transverse*, or *oblique*. In the case of domical, nearly isometric folded structures, one may speak of *radial*, *concentric*, or *oblique* joints.

A somewhat different classification must be used for massive rocks (Fig. 41). Here the position of the joints will be determined primarily in relation to the planar (parallel to the surface) or linear orientation of the internal structure of the rock. Thus longitudinal and transverse joints will be those which run, respectively, with or against the “grain” of the rock. Sometimes diagonal joints are also distinguished. In relation to the shape of the massive rock, the following types of joints may be distinguished: (1) *bordering* joints parallel to the outer surfaces of

the massive rock; (2) *radial* joints, perpendicular to the same surfaces; (3) *surface* joints, which usually repeat the surface of the earth in more leveled form within the massive rock; and (4) *enclosed* joints, which cut out rectangular and isometric (at least in one section) divisions of the rock, forming columnar, spherical, and pillow structures, etc. Enclosed joints occur mainly in extrusive but sometimes also in intrusive rocks.

Joints usually form a *set*, or group of parallel joints. As a rule, one encounters a number of sets intersecting at various angles and thus forming a *joint system*, which cuts the rock into blocks of one shape or another. Such blocks are called *joint blocks*. Their shapes depend on the directions of the various joint sets and their frequency in the various directions. The frequency of the joints will vary greatly in different sets. It may be expressed as the average number of joints per meter in the direction perpendicular to the given set of joints.

A special case is that of the frequently encountered sets of joints en echelon, as shown in Fig. 42. Here there is a zone containing a series of

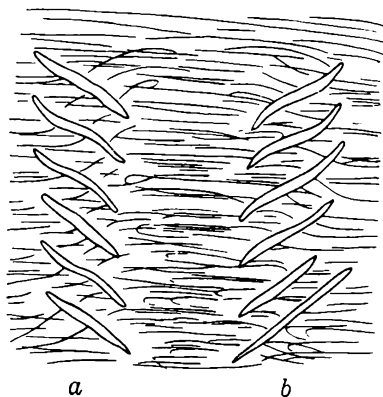


FIG. 42. En echelon joints: (a) left, (b) right.



FIG. 43. Branching joints (sketch).

parallel, relatively short joints whose strikes are at an angle to the trend of the zone. Such sets may be divided into right and left echelons. In the left set or echelon (Fig. 42a) the observer, looking in the direction of the zone, sees the joints moving off to the left; the reverse is true in the right set or echelon (Fig. 42b). Sometimes joints are bent, so that a single joint, in bending, may pass from one set into another.

Branching joints are of considerable interest. A cluster of branching joints will often form a so-called horsetail structure (Fig. 43). Sometimes joints, after branching out, come together again.

The classification of tectonic joints and joint sets given above is purely morphologic and may be used in brief descriptions of the observed features of joints and groups of joints. Another classification may be offered, however, based on a reconstruction of the mechanical conditions under which the joints were formed. This classification will be referred to again in discussing the mechanism of tectonic joint formation.

FAULTS

The richest material on the morphology of faults has been obtained from detailed studies of folded coal basins now being worked. Such basins usually contain a large number of small and large faults, which must be studied in order to survey and work the basin properly and, above all, to develop the most effective methods of searching for displaced seams. The tracing of beds and fault planes in underground workings makes it possible to carry out such studies with great accuracy.

It has already been mentioned that a great contribution to the geology and geometry of tectonic faults has been made by the geologists who studied the Kuznets coal basin, and particularly by M. A. Usov and his students (Usov, 1940; Belitskiy, 1949). The discussion that follows is based primarily on their material, although, as will be seen, in a number of essential points exceptions will be taken to Usov's classification of faults.

The characteristics of simple faults will be considered first. Simple faults are divided into *separations*, *strike-slip faults*, *normal faults*, *reverse faults*, and *thrusts*. In addition, the term *underthrust* will be used tentatively.

SEPARATIONS

A separation is the displacement produced by simple parting of the walls of the fracture. As a result of separation, the opening of the fracture is increased. Separations are most frequently observed in relatively

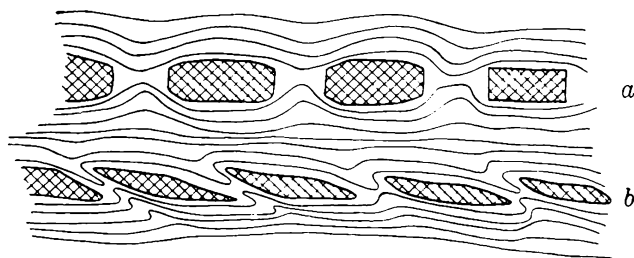


FIG. 44. Two types of separation: (a) breaking, (b) shearing.

competent beds lying between more plastic beds, when the entire suite of rocks is subjected to pressure normal to the bedding and to tension parallel to the beds. Under these circumstances the plastic rocks flow and change their shapes but maintain their coherence, and the more brittle beds between them are ruptured into separate blocks, which are then moved apart. The more plastic rock usually flows into the spaces

between them, but sometimes these interstices are filled with exogenic mineral matter carried in by ground waters (Fig. 44).*

Separations can vary in amplitude: in some cases they may attain several scores of meters, but more often they do not exceed several meters.

STRIKE-SLIP FAULTS

Strike-slip faults are faults in which the movement is along the strike of the plane of fracture. The dip of the fracture or fault plane may be vertical or inclined at any angle. Right and left strike-slip faults are distinguished by the direction of the displacement: if the fault is seen from the side, in the section perpendicular to the fracture, the more distant block from the observer is the one displaced to the right in a right strike-slip fault. In a left fault, the more distant block is displaced to the left.

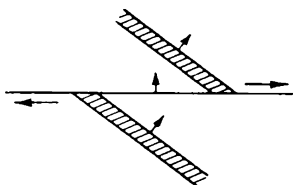


FIG. 45. Right oblique positive con-formable strike-slip fault (map).

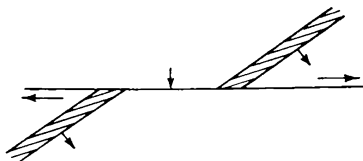


FIG. 46. Right oblique negative non-conformable strike-slip fault (map).

In nonlayered rocks the strike-slip fault is characterized only by its right or left relative displacement, the horizontal amplitude of this displacement, and the position of the fault plane in space. A strike-slip fault that cuts through a layered rock composed of horizontally lying beds has much more complicated characteristics. According to the relationship between the strike of the fault plane and the strike of the beds, one may distinguish *longitudinal*, *oblique*, and *transverse* faults. A longitudinal strike-slip fault results in no visible displacement either in the horizontal or the vertical section. In the case of oblique faults the geometry of the displacement will vary. Figure 45 shows a right strike-slip fault, with displacement in the direction of the obtuse angle between the beds and the fault plane, when seen in the horizontal projection. This will be called a *positive* strike-slip fault; it is associated with repetition of the beds. Figure 46 shows a right strike-slip fault

* EDITOR'S NOTE: Rupture of this kind is usually termed *boudinage* (see Marland P. Billings, *Structural Geology*, 2d ed., 1954, p. 354). Beloussov, in his discussion of distortion in Chap. 6, uses "boudinage" to refer to the squeezing of a more competent bed into tectonic lenses without actual rupture between the lenses.

with displacement in the direction of the acute angle between the beds and the fault plane; this is a *negative* strike-slip fault, characterized by the omission of beds.

Both positive and negative strike-slip faults may be either *conformable* or *nonconformable*. Conformable faults are those whose fault planes dip in the same direction as do the beds, whereas in nonconformable faults the fault plane and the beds dip in opposite directions. Such a distinction is possible, obviously, only in the case of an inclined fault plane and an oblique or longitudinal fault and cannot be made for transverse strike-slip faults. Transverse strike-slip faults can be only right and left, vertical and inclined.

Thus strike-slip faults may be right or left, inclined or vertical, longitudinal, oblique, or transverse, positive or negative, and conformable or nonconformable.

GENERAL OBSERVATIONS ON NORMAL FAULTS, REVERSE FAULTS, AND THRUSTS

In the case of normal faults, reverse faults, and thrusts the displacement is in the vertical plane perpendicular to the plane of the fault. The difference between reverse faults and thrusts, however, is not always accurately defined.

According to accepted terminology, an *ideal normal fault* involves downward movement of the hanging wall relative to the foot wall. It would be more proper to call this a *common normal fault*. In a classification based entirely on geometry the *reverse fault*, or relative upward movement of the hanging wall, should in all cases be called either an upthrust or overthrust. When there are none but geometric criteria for distinguishing various faults, it is recommended that all relative movements of the hanging wall downward be called *normal faults* and upward movements be called *upthrusts* and *overthrusts*. It then remains only to agree on a term for displacements along a vertical fracture plane and to draw the line between upthrusts and overthrusts. Vertical displacements may be grouped with normal faults, and reverse faults in which the fault plane dips more steeply than 60° may be called upthrusts, whereas more gently dipping upward displacements may be called overthrusts.

If, however, the classification of faults can be based not only on the morphology of the structure but also on the general character of the actual movement of the earth's crust that has resulted in the fault, the grouping of faults in one category or another must have a more solid foundation. In such a case, it is more accurate to call a normal fault one that has been produced by a *downward* movement of one part of the earth's crust *relative* to another. Such a normal fault might then be

common, vertical, or reversed. On the other hand, the displacement associated with a relative *upward* movement of a part of the earth's crust in comparison to another should be called an upthrust, regardless of whether the hanging wall or the foot wall has moved upward or whether the movement was along a vertical plane. In this classification a fault produced by a horizontal movement or tangential displacement of the earth's crust will be called an overthrust.

It would seem at first glance that with such an interpretation of the terms it is completely impossible to distinguish between normal faults and upthrusts, since vertical movements of individual parts of the crust are always relative and one cannot determine whether the dislocation in question is associated with absolute upward or absolute downward movement. This danger of confusion is, actually, present in many cases. If, for example, a dome is changed by numerous faults into a "broken plate" and divided into elevated and depressed sections, one cannot distinguish between normal faults and upthrusts in this sense. But cases are frequently encountered in which it is possible to establish whether the fault was produced by upward or downward absolute movement. An isolated tectonic upwarp, for example, such as an elongated uplift or a dome located in an extensive region of almost horizontally lying beds and fractured by a fault produced by the upward movement, may obviously be included in the category of upthrusts at least in the genetic if not in the geometric sense.

In similar regions of horizontally lying beds one finds grabens bordered on both sides by faults that should, from this point of view, be called normal faults. In both the cases cited, the attribution of a fault to one type or another does not depend on the inclination of the fault plane or the relative displacement of the foot wall and the hanging wall. The criterion for calling a fault an overthrust should be its genetic association with continuous folding.

Thus the expressions "normal fault," "upthrust," and "overthrust" must be understood in two senses. The first, purely morphologic geometric meaning must be used when the general nature of the movement of the earth's crust that has produced the fault is unknown or unclear. In the morphologic classification, normal faults, upthrusts, and overthrusts do not conflict in meaning: the movement sense of normal faults is opposite to that of upthrusts and overthrusts, which are always reverse; upthrusts are steep (over 60°), and overthrusts are always gently sloping.

In the second place, the terms "normal fault," "upthrust," and "overthrust" may be used in a genetic sense—that is, taking account of the conditions under which the fault was formed and the general nature of the movement of the earth's crust. In this classification a normal fault is always linked with a downward movement of a portion of the earth's

crust, an upthrust with an upward movement, and an overthrust with a horizontal displacement usually associated with folding.

The attempt should constantly be made to use the terms in the second sense, by studying all the tectonic circumstances under which a given fault was produced. In the second classification the meanings of "normal fault," "upthrust," and "overthrust" may overlap, in the morphologic sense, since structures of the same form may be produced by different tectonic movements of the earth's crust.

The individual types of faults will be described below in more detail.

NORMAL FAULTS

The normal fault (in the morphologic sense) is a displacement along the dip of a fracture with the hanging wall moved downward, or a downward displacement along a vertical fracture plane. Thus the normal fault may be inclined or vertical. The normal fault in layered rocks

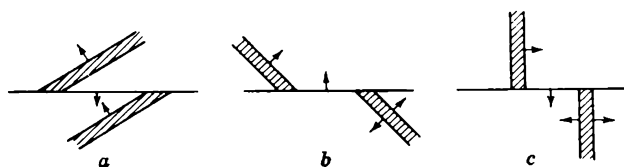


FIG. 47. Straight normal faults (map): (a) oblique nonconformable; (b) oblique conformable; (c) transverse.

whose beds are inclined to the horizontal has a number of varieties, depending on the relative positions in space of the beds and the fault plane.

Depending on the strike of the fault plane, faults may be *longitudinal*, *oblique*, or *transverse*. In longitudinal faults the plane is parallel to the strike of the beds, in oblique faults it intersects the bed at any angle to the strike between 0° and 90° , and in transverse faults it is normal to the strike of the bedding.

Longitudinal and oblique normal faults may be *conformable* or *nonconformable*, depending on the inclination of the fault plane. Conformable faults are longitudinal or oblique normal faults whose fault plane dips in the same direction as the beds they intersect; the reverse is true of nonconformable normal faults (Fig. 47).

Certain geometric results of displacement along normal conformable and nonconformable faults will be considered below. If the fault plane dips more steeply than the bed, a straight conformable normal fault, whatever the position of the fault plane and in any vertical section, will result in a gap between the displaced parts of the bed. A line perpendicular to the bed and passing through the fault will not encounter

the continuation of the bed. In the horizontal section (on a map) the hanging wall is drawn back (against the direction of the dip). When the fault plane of a conformable normal fault dips more gently than the bed, the beds are repeated. When the fault plane of a nonconformable normal fault grades from a vertical dip to one perpendicular to the beds, the displacement results in a repetition of beds; otherwise, there is an omission of beds.

If one abandons the purely geometric meaning of the term and gives it a genetic meaning, it becomes evident that the most common variety of normal fault has a downward movement of the hanging wall. This is due to the fact that downward movement of part of the earth's crust in normal faulting is usually the result of subsidence caused by extension. Such extension may occur, for example, on the crest of an upwarp (see below); one would expect normal faulting to accompany extension of part of the earth's crust.

But a downward displacement of the foot wall also occurs as a particular manifestation of normal faulting; such faults are called *reversed normal faults*. Morphologically they are identical with common upthrusts, and thus it is more convenient to consider their geometric features below, in the section devoted to upthrusts. Common and reversed normal faults are closely associated and usually are part of a single structure. It is often observed, for instance, that the plane of a normal fault is wavy and that its dip changes, dipping sometimes in the direction of the downthrown and sometimes in that of the upthrown wall. In addition, the direction of the relative displacement of the walls may change, so in some places one wall and in other places the other wall is the downthrown block. Such faults are called *hinge faults*.

Thus a normal fault, in the genetic sense of the word, may be common or reversed, vertical or inclined, longitudinal, oblique, or transverse, and conformable or nonconformable. A change in the direction of relative displacement of the walls produces a hinge fault.

UPTHRUSTS

It was seen above that, in the morphologic sense, an upthrust is the same as a reversed normal fault. The genetic concept, however, broadens the meaning of upthrust to include all the faulted structures produced by upward movement of a portion of the earth's crust relative to another part.

A common upthrust is one in which the hanging wall has been displaced upward; this includes faults with vertical fault planes. Depending on the position of the fault plane in relation to the beds cut by the fault, as in the case of the other faults, one may distinguish longitudinal, oblique, and transverse, conformable and nonconformable upthrusts. A

common conformable upthrust, which has the same shape as a reversed conformable normal fault, results in repetition of beds if the fault plane dips more steeply than the beds and in omission if it dips more gently. A common nonconformable upthrust produces omission when the fault plane lies between the vertical and the line perpendicular to the beds; at other angles it causes repetition.

If the foot wall of an upthrust (in the genetic sense) moves upward, a structure is produced that is morphologically the same as a common normal fault. This is a reversed upthrust, which can be conformable or nonconformable. The geometric result of displacement of the beds in a reversed upthrust is the same as that of a common normal fault.

Thus upthrusts may be common or reversed, longitudinal, oblique, or transverse, and conformable or nonconformable.

OVERTHRUSTS

An overthrust is a fault arising mainly as a result of horizontal tectonic movement—in other words, of tangential compression. In the

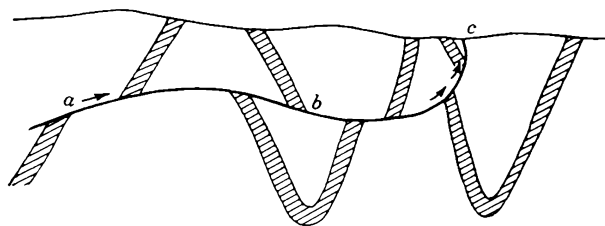


FIG. 48. Overthrust: (part *a*) common; (part *b*) plunging; (part *c*) reversed.

structure of the earth's crust, overthrusts are very closely associated with continuous folding: they are, in fact, a further development of continuous folding.

As in the types described above, overthrusts may be longitudinal, oblique, or transverse. Since the fracture planes of overthrusts are more curved and wavy than those of other faults, these definitions may refer either to the segment of the overthrust or to its mean direction over a smaller or greater extent.

A common overthrust is one in which, as in the case of upthrusts, the hanging wall is displaced upward (Fig. 48*a*). Extremely large overthrusts sometimes have gently sloping and undulating thrust planes, which, because of the undulations, may in places have a reversed slope. In addition, there are sections in which the overthrust must be said to be plunging downward (Fig. 48*b*). Finally, parts of the overthrust may have an overturned thrust plane (reversed overthrust, Fig. 48*c*). Plunging and reversed overthrusts are morphologically the same as a common

normal fault. If the details of the movement in the immediate vicinity of the fault are analyzed without regard to the structure as a whole, a reversed overthrust may be confused with a reversed upthrust.

Depending on the position of the fault plane in relation to the attitude of the beds, as in the preceding cases, one may distinguish conformable and nonconformable overthrusts. The resulting repetitions and omissions of beds should be clear without further explanation (see Fig. 48).

The nomenclature for overthrusts proposed here is closer to familiar geologic terminology than M. A. Usov's classification, which calls the common overthrusts described here reversed, and vice versa. Overthrusts may thus also be common or reversed, longitudinal, oblique, or transverse, and conformable or nonconformable.

UNDERTHROWS AND UNDERTHRUSTS

M. A. Usov (1940), in his classification of faults, also distinguished underthrows and underthrusts, meaning upthrusts and overthrusts produced by active movement, not of the hanging wall, but of the foot

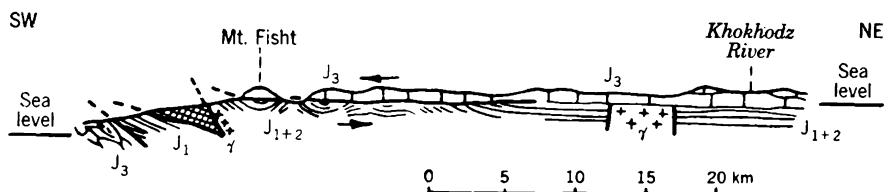


FIG. 49. Underthrust in the Mt. Fisht area of the northern Caucasus (after V. V. Beloussov).

wall of the fault. In the terminology used in this book, "underthrow" is replaced by "reversed upthrusts." The term "underthrust" may in some cases be retained, if it is clear from the total tectonic environment that the structure in question is the result of active movement of an actually underthrust foot wall and not of an overthrust. A structural section of one of the areas of the northern Caucasus shows an example of a fault that may be properly called an underthrust: here the shale series of the Lower and Middle Jurassic were thrust northward under immovable Upper Jurassic limestones (Fig. 49). Nevertheless, the term "underthrust" will be used only in rare cases.

Table 10 is a compilation of common, reversed, and plunging normal faults and upthrusts and overthrusts, showing the varieties of these faults that are morphologically identical.

It might appear that the use of terms in two senses—morphologic and genetic—would lead to misunderstanding, since any term will have one meaning at one time and the other meaning at another. Actually, how-

ever, this does not happen. One must attempt, of course, to use the genetic term in all cases. But if this is impossible because of lack of data, or if different investigators produce different interpretations of the origin of any given fault, there will be no danger of inconsistency in terminology, since the morphologic meaning of each term is fully and accurately defined, so long as the term for the variety (common, reversed, etc.) is always used along with the basic word for the dislocation (normal fault, upthrust, overthrust). If a fault is called a normal fault by one geologist and an upthrust by another, there could, of

Table 10. Varieties of Faults

Normal fault...	Upthrust	Overthrust	Normal fault	Upthrust	Overthrust
Common.....	Reversed	Plunging, reversed	Reversed	Common	Common

course, be a misunderstanding; but if one geologist speaks of a common normal fault and the other of a reversed upthrust, it is obvious that they are both describing the same structure. The morphology is perfectly clear, and the difference in terminology is due to different interpretations of the origin of the given fault.

COMPLEX FAULTS

The characteristic features of simple faults, in which the displacement follows either the strike or the dip, have been given above. As also noted above, however, such simple displacements are a rarity in nature. It is true that for ordinary purposes, when particular accuracy is not required, some deviations of the displacement from the dip or strike direction can be disregarded. But the majority of displacements through faulting that are observed in the field belong to the category of complex faults, in which movement along the dip (the vertical component) is combined with movement along the strike (the horizontal component), so that the resulting displacement is in an oblique direction.

The geometric analysis of such complex displacements in the field is sometimes an extremely difficult but nevertheless exceedingly important practical task (Molchanov, 1935; Belitskiy, 1949; Bauman, 1907). The most appropriate names for complex displacements are those which most accurately reflect the compound nature of the movement. Thus, if the downward displacement of the hanging wall is not directly along the strike of the fault plane but is oblique to it, one must speak of strike-slip normal faults or normal strike-slip faults. The difference between these is that the first (strike-slip normal fault) is produced by movement that is more downward than to the side, or closer to the dip than to the strike, whereas movement along the strike predominates in

the second case. In just the same way one can distinguish strike-slip overthrusts and overthrust strike-slip faults, upthrust strike-slip faults and strike-slip upthrusts. The second term indicates the predominant movement.

As in the case of simple faults, the position of the fault plane relative to the attitude of the beds is the basis for distinguishing longitudinal, oblique, and transverse faults, on the one hand, and conformable and nonconformable strike-slip normal faults, overthrust strike-slip faults, strike-slip upthrusts, etc., on the other hand. In describing the displacement, it is also important to know the direction of the strike-slip component—along the dip of the beds or against the dip.

The geometry of complex displacements will not be examined in detail here; this is the task of structural geometry. This discussion will

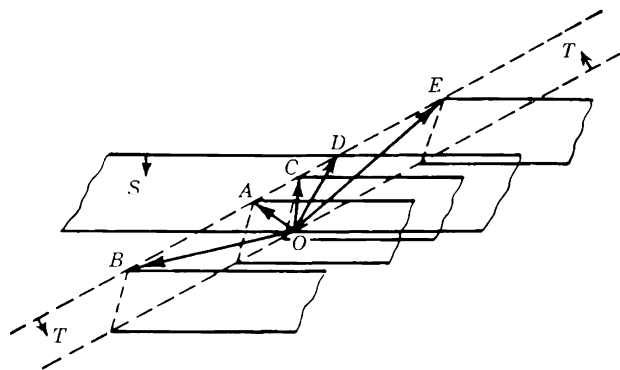


FIG. 50. Diagram of complex conformable strike-slip normal faults and upthrust strike-slip faults (after A. Belitskiy).

limit itself to one example, which will give some idea of the peculiar difficulties involved in the interpretation and description of these displacements. Figure 50 shows a series of multiple strike-slip normal faults and upthrust strike-slip displacements. The plane *S* is the bedding plane, which dips toward the reader; the plane *T* is the fault plane, which cuts the bed at an acute angle and also dips toward the reader. Hence this is a conformable fault. If, as a result of the displacement, point *O* moves to point *A* along the fault plane, it is a straight conformable upthrust. Figure 51a shows a section (along the line I-II) and a map of this upthrust, with the resulting repetition of beds. If the displacement of point *O* follows the line *OB* (see Fig. 50), the fault is an upthrust strike-slip fault, because the strike-slip displacement here predominates over the upthrust. The section and map of this dislocation (Fig. 51b), like those of the previous one, show a repetition of beds. Repetition also results from the strike-slip upthrust along the line *OC* (see Figs. 50 and 51c). When the displacement is in the direction *OD*

(see Fig. 50)—that is, parallel to the bedding—there is no visible result of movement either in the section or on the map (see Fig. 51*d*). The most interesting case is that of displacement along the line *OE* (Fig. 50). In this upthrust strike-slip fault, as can be seen from Fig. 51*e*, there is an omission of beds both in the section and on the map. The unique feature is that the left wall, which appears from the section to have moved down, has actually moved upward. To understand this seeming

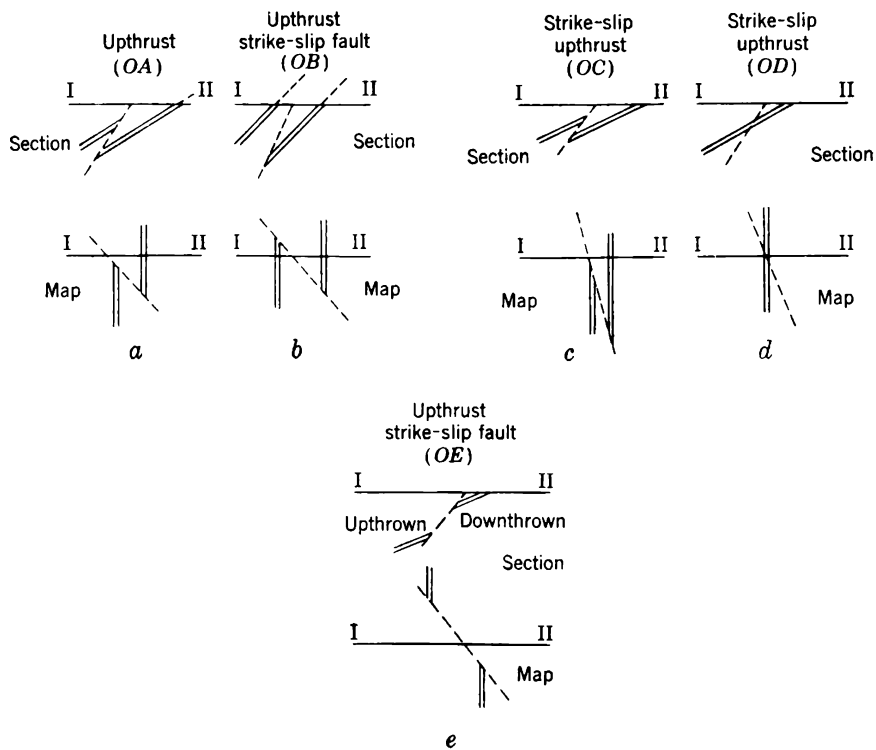


FIG. 51. Maps and sections of various complex faults.

discrepancy, it must be remembered that the movement of the left wall (see Fig. 51*e*, section) is directed mainly toward the reader and partly upward along the fault. But the upward movement is compensated and exceeded by the dip of the bed away from the reader. Since the left wall has been elevated along the fault plane at a smaller angle than the dip of the bed (projected on the fault plane), the line of intersection with the plane of the section moves downward during the displacement.

Similar combinations are possible for other complex faults. Thus it becomes clear that one must be very careful in judging the relative displacements of the walls of a fault from sections and maps. These will give a clear idea of the character of simple displacements, but in the

case of compound displacements and various combinations of the relative positions of the fault plane and the bedding plane, sections and maps are far from easy to interpret and hasty conclusions may lead to mistakes.

Two more simple examples will be given. Figure 52*a* shows a map of a structure containing two faults. If the attitudes of the fault plane and the beds are disregarded, one might conclude that this structure has been produced by strike-slip displacements associated with compression along the x axis and extension along the y axis. But if the dips of the fault plane and the beds are considered and if it is assumed that the

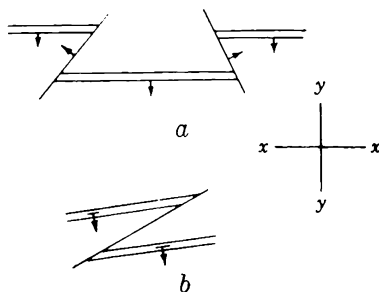


FIG. 52. Complex displacements: (a) maps; (b) section.

vertical (normal fault) component has been predominant in these displacements, the same structure will be seen as the result of elongation along the x axis and shortening along the y axis.

Figure 52*b* shows a section of a simple structure, which will immediately be interpreted as an overthrust of the left wall over the right wall and, consequently, as contraction along the x axis and extension along the y axis. But if the bed dips steeply toward the reader and the displacement of the left wall is also primarily toward the reader, with a fairly gentle downward movement along the fault plane, the same section will appear as the result of lowering of the left wall relative to the right, and the cause of the structure will be extension along the y axis and contraction along the x axis.

CHAPTER 8

TECTONIC CHANGES IN THE INTERNAL STRUCTURE OF ROCKS

PRIMARY INTERNAL STRUCTURE (TEXTURES) OF ROCKS

The dislocations discussed thus far have all been changes in the structures of rocks and in the external shapes of the geologic bodies that they compose. But tectonic forces change not only the occurrence of rocks—their external structures—but also the internal structure of the rocks themselves. These changes to a great degree consist in redistribution of the minerals or grains within the rock and in fracturing of the latter. If changes in the structural occurrence of rocks may be called *external tectonic dislocations*, the changes in their internal structure can be termed *internal tectonic dislocations*. The combination may be called *dynamic metamorphism*.

Just as the final morphologic result of external dislocation depends on the primary structure of the rock, the final result of internal dislocation depends not only on the character of the deformation and the mechanical properties of the rock but also on the primary internal structure. Very rarely do rocks have an amorphous, completely homogeneous internal structure; this is true for the most part only of volcanic glass, which occurs in young volcanogenic strata and is generally not prominent in geologic sections. The vast majority of rocks have a heterogeneous internal fabric characterized by a clastic or crystalline structure. In the first case the rocks are composed of fragments of other rocks or of minerals; in the second case they consist of crystals of different composition and shape.

The internal structure of a rock is determined by the shapes and dispositions of the grains or crystals in it. These elements of the internal structure of rocks are usually designated by the term “texture,” which will be used henceforth.

In sedimentary rocks the primary texture is determined by the conditions under which the sedimentary material was deposited. One of the most important factors is the movement of the medium in which the deposition occurs. On the bottom of the sea the movement of the water is primarily horizontal, because of the oscillations of the waves, which

are reflected at the bottom in a constant movement of water particles toward and away from the shore. Consequently, the particles of sediment are rolled back and forth along the bottom, finally coming to rest in the position most suitable to them under these circumstances. In the case of flat and elongated particles, the most suitable position is horizontal. For this reason the primary texture of marine sediments is characterized by the disposition of elongated and flat grains parallel to the bedding. In the zone exposed to the action of the surf, the arrangement of the particles frequently becomes more complex: here the grains lie in such a way as to form a rippled surface. Such textures are very rarely preserved in geologic sections, however, since shore-line deposits are to a considerable degree eroded away during the advance and retreat of the sea.

Such foliated primary texture is particularly clear in argillaceous deposits, in which it is determined by the horizontal position of the tiny parallel flakes of sericite, mica, and other minerals. Such texture is sometimes also noticeable in sandy deposits, if the grains are of elongated shape. It is almost always clearly evident in conglomerates, inasmuch as the pebbles usually have the shapes of more or less elongated and simplified ellipsoids. The horizontal texture of limestones is due to the flat-lying position of the organic remains that compose them. In strongly recrystallized limestones, on the other hand, such texture may be to a considerable degree veiled and replaced by a massive texture with a random disposition of calcite crystals. In sediments of chemical origin the individual grains are frequently eroded and redistributed along the bottom, thus acquiring a horizontal orientation; often, however, one may observe a massive primary texture in such rocks. The creation of a primary horizontal texture in sedimentary rocks is facilitated by seasonal and, in general, climatic changes in the composition of the sedimentary material, leading to the formation of fine primary bedding in the deposits.

The primary texture in sediments of continental origin is usually more complicated. Here, among the deposits of fluvial and aeolian formation, there is widespread cross-bedding of various forms, with the grains lying at an angle to the bedding plane. In stream channels the grains, in the form of imbricated overlapping lamellae, are ordinarily not horizontal but dip upstream.

Primary textures in *igneous rocks* show greater variety and complexity than those in sedimentary rocks (Pek, 1939; Balk, 1937). In extrusive rocks—lava flows—the primary texture is determined by the flow of the lava and characterized by the disposition of crystalline and gaseous inclusions in streams along the flow lines. This creates *fluidal texture*. Since the flow is usually complex, with twists, bends, and turns, the fluidal texture also contains wavy and loop-shaped bends, etc.

In exactly the same manner, the subterranean, rather than surface, flow of the magma determines the primary texture of intrusive rocks. This texture depends to a considerable degree on whether the magma was crystallized after or during the flow. In the first case, the crystallization of the igneous melt will take place under quiescent conditions, in a nonmoving mass, so that the crystalline texture shows almost no orientation and the individual crystals are variously disposed. This is the primary structure of the rocks of which many batholiths are composed. These rocks as a rule show no predominant orientation in their crystalline texture. This is evidence that the magma crystallized after the movement had come to an end.

Primary textures formed by crystallization simultaneous with the movement of the magma are most clearly seen in the igneous bodies that have been classed as "minor" intrusives: laccoliths, stocks, drop-shaped intrusives, etc. They are also seen in fissure intrusives and sills.

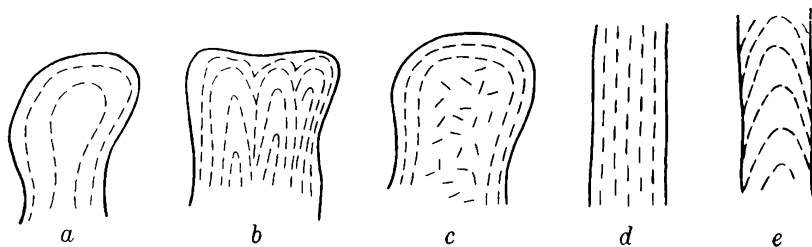


FIG. 53. Primary oriented textures of intrusives: (a) domical texture; (b) complex domical texture; (c) combination of oriented texture on the periphery and non-oriented texture within the intrusive; (d) longitudinal texture in a dike; (e) bent texture in a dike.

In every case these textures are reflected in the oriented disposition of the flat and elongated crystals in the rock, as well as in the orientation of the schlieren. The dispositions of these elements form various regular figures that are closely connected with the shape of the intrusive.

Two varieties of oriented textures are usually distinguished: *parallel planar*, or simple *planar*, and *parallel linear*, or simple *linear*. The first term designates the parallel disposition of the flat minerals (such as mica, chlorite, and epidote). Linear texture takes the form of a parallel disposition of the long axes of minerals, which may have their flat surfaces (if any) turned in all directions about these long axes. Planar and linear textures are frequently observed together; in such a case, both the flat sides and the long axes of the minerals are parallel. In parallel planar texture that is not linear, the flat sides of the minerals are parallel to each other, but the long axes may point in various directions within the parallel planes.

The figures formed by the oriented positions of the minerals within

the intrusives may be extremely varied (Fig. 53). Dome-shaped textures, which follow the lines of the usually domical top of the intrusive, are very often seen. In the simplest case, the oriented minerals are everywhere parallel to the top and the sides of the intrusive (Fig. 53a). But complex domical oriented textures are also encountered, in which the interior of the intrusive appears to be made up of several domes (Fig. 53b). Not infrequently the domical texture occurs only in the outer shell of the intrusive, and the deeper parts of it show no orientation of the crystals (Fig. 53c). Fissure intrusives have oriented textures of two types: either parallel to the walls of the fissure (Fig. 53d) or forming convex meniscuslike bends in a single direction (Fig. 53e).

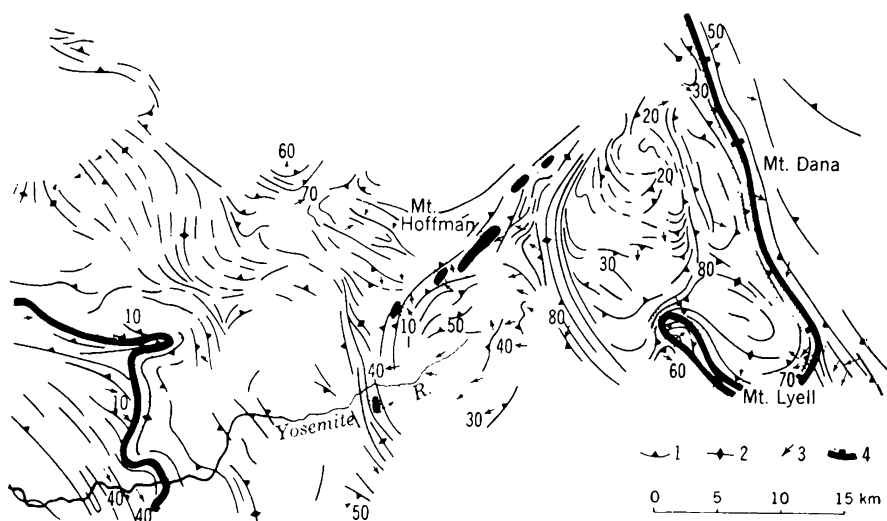


FIG. 54. Map of flow structure in the Yosemite intrusive in California (after H. Cloos and R. Balk): (1) strikes and dips of flow foliation; (2) vertical foliation; (3) linear flow structure; (4) contact with host rock.

In determining the conditions and mechanism of the formation of an intrusive, it is extremely important to study the oriented textures in it. The results of such a study are plotted in sections or maps. Figure 54 shows a map of an intrusive with indications of the oriented textures in various parts of its surface.

THE SECONDARY TEXTURE OF ROCKS

The textures in rocks that are formed by tectonic deformation very closely resemble many primary textures. For this reason, it is far from simple to distinguish between primary and secondary, or dislocational, textures. This applies mainly to plastic internal dislocations that lead to

the formation of oriented textures similar to primary ones. It is usually easier to recognize internal dislocations accompanied by rupture, as in the brecciation of a rock, although doubt sometimes arises as to whether the breccia in question is primary, and of sedimentary, or secondary, and of tectonic, origin.

Internal tectonic dislocations always appear in connection with external dislocations; this circumstance facilitates their recognition. In different places, depending on the intensity of the external dislocation and the mechanical properties of the rocks under the given conditions, the internal dislocations are manifested in very different forms. In some places, the internal dislocations encompass very large volumes of the rock, wherever all the rocks have undergone dynamic metamorphism to some degree. In other places, the internal dislocations are associated with individual parts or zones connected with specific external dislocations, which are most often faults. Thus one may speak of *local* and *general* dynamic metamorphism. In the overwhelming majority of cases, the disjunctive internal dislocations are local; cohesive internal dislocations are frequently general or regional. Rocks that have undergone internal dislocations and have a corresponding secondary texture are called *tectonites*.

We shall, first of all, discuss the cohesive internal dislocations manifested in oriented textures (Pek, 1939; Balk, 1937; Grubenmann and Niggli, 1933; Cloos, 1946; Sander, 1948–1950). Even though a sedimentary rock does not have a primary oriented texture, such an oriented texture may appear as the result of tectonic action. If a primary oriented texture exists and the strains are in the same direction, such texture is underscored and intensified or else is rearranged to correspond to the new direction. Hence the most doubtful case is that in which the oriented texture is parallel to the bedding: under these circumstances it may be either primary or secondary.

Oriented texture parallel to the bedding planes is sometimes called *bedding cleavage*, or *bedding schistosity*; it is very widespread in ancient (Archean) metamorphic and gneissic strata. The fact that in such rocks the oriented textures may be only very remotely connected with primary texture and fundamentally determined by tectonic deformations is confirmed by the intensive general dislocation of the gneisses and their recrystallization, in the course of which the primary shapes of the grains in the original sedimentary rocks have completely disappeared (without the disappearance, however, of the oriented texture), by the very distinct orientation of the crystals (much more distinct than that usually observed in sedimentary rocks), by the clear change in the shapes of the residual sedimentary features that are sometimes preserved in these rocks, and by the deformation of the crystal lattice of the minerals, as determined by studies under the microscope.

Pebbles sometimes occur as relict sedimentary structures in metamorphic rocks and may be seen on weathered surfaces. These pebbles as a rule have distorted shapes: they are strongly flattened and broken up, with elongated ends ("caudate" pebbles). Their entire shape testifies to their being crushed by tectonic forces. The flattening of the pebbles follows the direction of the oriented texture in the surrounding rocks (Fig. 55). Such secondary flattening is also seen in sand grains, whose longitudinal section has a flattened or lenticular shape.

A very clear picture of the deformation of minerals is seen in investigations with the microscope (Sander, 1948-1950). With polarized light under the microscope one may distinguish in thin sections indications of deformation of the crystal lattices of the minerals (wavelike extinction, etc.). Moreover, the oriented shapes of the crystals are very clear under the microscope (Fig. 56).

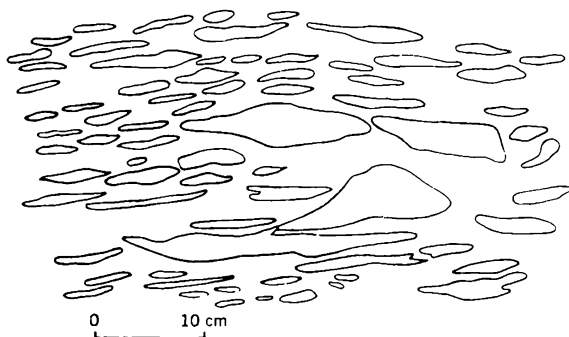


FIG. 55. Crushed and flattened ("caudate") pebbles.

When there is a clear parallel-oriented texture, the rock is easily broken up into thin plates corresponding to the planes of orientation.

In younger and less metamorphosed but nevertheless strongly deformed sedimentary rocks, the oriented textures frequently intersect the bedding at some angle. This is the *flow cleavage*, developed in sedimentary rocks that have been compressed into folds. At the surfaces of outcrops this cleavage takes the form of a very fine shearing of the rocks. Study of this cleavage under the microscope reveals that it is most intensively manifested in shales, although it also appears in other rocks as well. When there is an alternation of beds with different compositions, it is often developed unevenly, primarily in the most plastic rocks, whereas in the harder rocks immediately adjacent (such as sandstones or limestones interbedded with shales) it may be absent.

Flow cleavage is most often parallel to the axial planes of the folds (*principal flow cleavage*). On the other hand, one often encounters so-called inverted fan cleavage, in which the cleavage surfaces of the

rocks converge upward like an upside-down fan. This is usually observed in thin beds of plastic rocks occurring among thick, less plastic rocks (for instance, in shales among thick sandstones). In addition, when the cleavage is localized in individual beds, it appears to be curved in the form of an S (S-shaped cleavage). Figure 57 shows the different manifestations of flow cleavage in folded strata.

In intrusive igneous rocks, secondary internal dislocations take the form of oriented textures in rocks that did not originally contain such



FIG. 56. Thin section of a rock that has undergone dynamic metamorphism.

textures and of the complication of the original flow patterns of rocks in which these existed earlier. The first phenomenon is observed in the transformation into gneiss, as a result of tectonic deformation, of granitic batholiths that originally had a massive uniform texture. Only after thorough petrologic study with the microscope, however, and then not always, can one tell whether the given rock is a massive primary gneiss or whether it acquired its oriented texture later, as a result of tectonic deformations.

In minor intrusives, which nearly always have a primary oriented texture, the secondary distortions almost invariably take the form of changes and complications in this orientation. In such cases it is ex-

tremely difficult to distinguish the secondary changes from the primary texture. Such a distinction, in fact, frequently loses all meaning, since the formation of minor intrusives (as well as batholiths) is simultaneous with the tectonic deformation of that entire part of the earth's crust, as a result of the basic tectonic movement. Under these conditions, the intrusive simultaneously undergoes internal and external deformation, so the distinction between the formation of primary and secondary textures is lost.

Let us consider disjunctive internal dislocations, which are primarily local, as already pointed out. The most widespread form of internal disjunctive dislocation is the fracturing of a rock associated with displacement along a fault, within the zone in which the fault affects the surrounding rocks (Grubenmann and Niggli, 1933). In such fracturing, the rock or the individual minerals in it are broken into pieces with

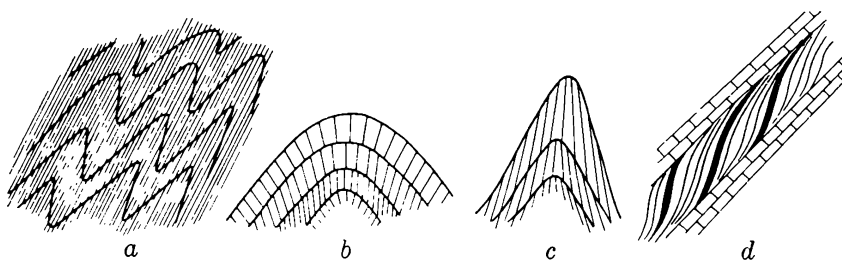


FIG. 57. Varieties of cleavage: (a) "principal" cleavage, parallel to the axial planes of the folds; (b) shear cleavage; (c) inverted fan-shaped cleavage; (d) S-shaped cleavage.

acute angles and irregular shapes, which are displaced from their original positions and moved about. Such fragments may vary greatly in size, from microscopic particles to great blocks tens or even hundreds of meters in cross section. Coarse fracturing that breaks the rock into large pieces easily seen with the naked eye produces *tectonic breccia*. The interstices between the large rock fragments in tectonic breccias are usually filled with more finely crushed particles of the same fractured rock—"rock flour" or "sand." Tectonic breccias are associated with both large- and small-scale displacements along faults; they are encountered along the surfaces of overthrusts, normal faults, strike-slip faults, and upthrusts.

The thickness of the brecciated rock depends on many factors; it is not true that a displacement of greater amplitude is necessarily accompanied by more intensive brecciation of the rock. Cases are known of faults with enormous displacements along the fault plane and little or no brecciation, whereas other, very small displacements may lead to the brecciation of a very thick series of rocks. Aside from the scale of the displacement and the composition of the rock, an important factor may

be the depth at which the dislocation occurs below the earth's surface; other conditions determining the mechanical properties of the rocks may also be important. The thickness of the tectonic breccia, in fact, may vary greatly along the line of a single fault.

A study of the transition between tectonic breccia and unfractured rock reveals that the brecciation passes through a stage corresponding to a so-called nutlike rock (*kakirite*); this is the initial stage of brecciation. This name is given to rocks shot through by multiple intersecting fractures (large and small) that divide it into separate fragments; these discrete fragments, however, are not yet displaced from their original positions in the rock. When struck with a hammer, this rock immediately falls into tiny angular fragments, so that it is often impossible to break off a piece with a fresh surface. A comparison of tectonic breccia with *kakirite* shows that the brecciation is preceded by the formation of a network of fractures associated with the given disjunctive dislocation or by a network of tiny secondary displacements accompanying the basic fault. As they gradually develop, these fractures break up the rock into a large number of blocks and pieces, which are then moved from their positions by the force of friction, turned over each other, and ground into the fissure formed by the thrust or fault.

Slickensides are formed on the surfaces of blocks and pieces that rub against each other. These slickensides often contain new minerals—the so-called stress minerals that are the result of recrystallization: quartz, epidote, sericite, chlorite, acicular hornblende, and chrysotile serpentine. These new minerals reflect the rock's adaptation to the new conditions. Great friction raises the temperature, which, in turn, increases the mobility of the ions. Even before the melting point is reached, the diffusion of the ions leads to their regrouping and the formation of new minerals, which because of their flat or acicular shapes produce smoother and more slippery surfaces and thus facilitate movement along these surfaces.

The interstices between the blocks, or the fractures in a *kakirite*, are frequently filled with secondary material, which may be carried in by water (such as calcite or barite) or may be of igneous origin. Hot solutions often deposit ore material in the fractures of a *kakirite* or in the interstices of a fault breccia. The wall rocks of many ore deposits are fault breccias. Moreover in rocks containing relatively easily soluble minerals, the fissures and interstices are often filled with the redeposited material of these minerals. A very common occurrence is that of veinlike, breccialike limestones; these are breccias cemented by redeposited material, but they are not always of tectonic origin.

Cataclasis is the name given to the formation of breccias whose fragments are of microscopic size. This process is manifested in the frac-

turing of the individual minerals, such fracturing may encompass all the minerals, thus turning the entire rock into a peculiar "tectonic sand." More often, however, the cataclasis is associated only with certain of the minerals, and especially with the boundaries between them. The first stage of cataclasis is the "wavelike extinction" of quartz, an indication of deformation of the crystal lattice.

Often a quartz crystal, seen under the microscope, appears to be relatively undisturbed in the center, which shows only a more or less clearly manifested wavelike extinction, whereas its periphery is composed of a rim of "crush quartz," fractured into tiny particles (Fig.



FIG. 58. Thin section of a rock that has undergone dynamic cataclasis.

58). This fracturing on the periphery indicates that the grains were rotated and rubbed against each other during the process of dislocation.

The formation of new minerals is observed to a greater degree in cataclastic rocks than in tectonic breccias. Such new minerals include calcite, quartz, epidote, and sericite. In rocks that have undergone intensive cataclasis, recrystallization and the formation of new minerals appear almost everywhere. Sericitization of alkaline feldspars, albitization of plagioclase with the formation of epidote, transformation of augite into hornblende, and chloritization of biotite are especially common.

Cataclasis develops under approximately the same conditions as brecciation, but it is more frequently associated with overthrusts than with normal faults, since the fine fracturing of the individual minerals requires very great general compression and finely disseminated differential movement within the rock. Cataclasis, however, is not associated only with the surfaces of displacements along faults. It often encompasses great masses of rock not connected with any particular surface of displacement and, in such cases, indicates a transition to general metamorphism of the rock as a whole. Such general cataclasis (very fine fracturing of the rock) is observed in certain strongly deformed rocks in which the internal movement, manifested as innumerable surfaces of slickensiding, reaches great intensity. This happens especially often in ancient rocks that have undergone later dislocations distinct from the dislocations to which they were subjected earlier.

The ultimate result of such tectonic fracturing of rocks is a *mylonite*. The mylonite usually contains evidence of very fine shattering or grinding of the grains, which are "rolled out" and acquire very much elongated and flattened lenticular shapes. Thus mylonites are a feature of plastic deformation and the varying degrees of fracturing that follow it.

This survey of internal disjunctive dislocations would be incomplete without some mention of *shear cleavage*, which is very common in folded rocks. Shear cleavage, as distinct from flow cleavage, is not associated with oriented textures. It takes the form of very frequent fractures in the rock, which divide it into fine plates 0.5 to 2.0 cm in thickness. Within the folds, shear cleavage is usually disposed in the form of a fan diverging upward in anticlines or else perpendicular to the beds (see Fig. 57*b*). Here the oriented texture may be totally lacking, or it may be in a direction different from that of the cleavage. In the latter case, the flow cleavage and the shear cleavage may intersect.

CHAPTER 9

STRUCTURAL COMPLEXES

Structures in the earth's crust never occur in isolation. They are always combined in various ways; their relationships are governed by certain laws, the discovery of which is an important problem in geotectonics.

Combinations of structures may develop in different ways. To begin with, they may be determined by a historical succession of stages in the formation of structures. A single part of the earth's crust may undergo repeated tectonic deformations, and the different stages of tectonic history produce different structures in the rocks. Thus the deformations are superimposed in time. This case henceforth will be called a *historical combination*.

In the second place, the same tectonic phase, under the same physical conditions and in the same area, may produce different structural forms. For example, compression of a certain part of the earth's crust may produce both cohesive and disruptive dislocations, depending upon the physical properties of the rocks involved. In addition, structures of different orders and dimensions may complicate each other. Large folds, for example, frequently are complicated by small ones in the course of tectonic development, but both large and small folds develop simultaneously as a result of the same deformation. This combination of structures will hereafter be called *mechanical combinations*.

Finally, structural forms may be combined regionally. There are certain regularities in the disposition of various structural forms and complexes in the earth's crust: these are *regional combinations*.

These three kinds of structural combinations will be examined below in detail.

HISTORICAL COMBINATION OF STRUCTURES

The historical combination of structural forms in the earth's crust can best be seen in the example of repeated folding. This combination takes the form of structural stages of different age separated by angular unconformities (Bogdanov, 1949).

The structural stages are determined by the fact that folding in the earth's crust, over the course of time, takes place during certain periods called *periods of diastrophism*. These are separated by calm periods in which there is no folding or in which the folding is very weak. A necessary condition for the formation of two or more structural stages is the deposition of new beds after the first folding ends or between the two periods of folding.

Figure 59*b* shows an angular unconformity between a lower series crumpled into folds and an upper one that is undisturbed. The lower series was deposited first and then folded (the folding might have begun during the deposition of the lower series). Later the area was elevated and the folded stratum partially eroded, until its surface became

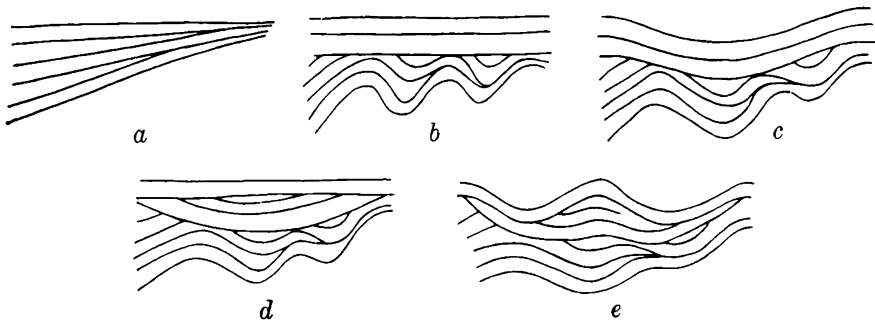


FIG. 59. Successive stages in the formation of a structure containing a number of angular unconformities.

smooth. Still later the earth's crust again subsided and was covered by new sediments, which form the horizontal layers of the upper series.

Figure 59*c* shows the result of the next tectonic stage, in which the earth's crust was again folded, forming broad upwarps and depressions. The upper series reflects only this later deformation, whereas the structure of the lower stratum is the result of both deformations.

Figure 59*d* shows the third structural stage, formed as the result of a new elevation and partial erosion of the earth's crust, and renewed subsidence and deposition of new flat-lying beds.

Figure 59*e*, finally, shows new folding involving all three series in this part of the earth's crust. In this case the upper series, the youngest one, has undergone only the latest stage of folding. The structure of the next lower series reflects both the latest and the penultimate folding, which produced the above broad upwarps and depressions. Finally, the lowermost series was involved in all three deformations.

It is worth dwelling in detail on the structure shown in Fig. 59*c*. The lower series demonstrates the typical continuous fold, but the next

higher series is deformed differently; the broad upwarps and depressions observed in it may be called *synclises* and *anticlises*. Thus structural forms of different types are found in different strata of the same area. The description of one occurrence of the rocks or another therefore cannot encompass the entire thickness of the earth's crust, but only certain more or less limited series. The overlying and underlying series may contain structures of entirely different types.

The above examples show that the simplest structural forms occur in a series of beds that has been deformed only once. Repeated tectonic dislocations produce superimposed deformations in the beds occurring at greater stratigraphic depths. To determine what structures existed before they were complicated by superimposed deformations, one must "take away" the results of later deformations and thus reconstruct the earlier structural history of the locality. The "removal," or subtraction, of later movements can be achieved, when there are a number of structural stages of different age, by bringing the upper strata into the horizontal position, one after another. In the structure shown in Fig. 59*e*, the uppermost series must be made horizontal, in imagination or in a drawing. Thus one obtains the structure shown in Fig. 59*d*, the structure that existed after the previous tectonic phase. By bringing the next lower horizon to the horizontal position one obtains Fig. 59*b*, whose structure corresponds to the earliest tectonic phase. When a higher structural stage is brought into a horizontal position, there must, of course, be an appropriate "correction" of the structure in the series below, which consequently acquires the "subtracted" structure, that produced by the previous folding. If one proceeds still further and straightens the top of the lowest series, until it becomes horizontal, the result will be the original sedimentary structure that existed prior to all the folding phases. Thus the primary structure of the beds will be regained; this is of great importance in reconstructing the history of the movements of the earth's crust. The peculiarities of this sedimentary structure are reflected, first of all, in lateral changes in the thicknesses of the deposits (Fig. 59*a*). The lateral distribution of various thicknesses and sedimentary facies yields a picture of the history of oscillatory movements of the earth's crust. Such movements could take place both before and during the folding. Bringing one structural stage after another into a horizontal position finally results in a reconstruction of the sedimentary conditions, regardless of folded structures. In studying the displacement of the primary thicknesses and facies, we are frequently forced to undertake this operation.

In addition to folded dislocations, the earth's crust also contains disruptive dislocations. In some cases, the same method of "removing" structural stages may be used to reconstruct their historical succession

(Fig. 60). Figure 60a shows a case in which a fault has equally disrupted both the upper and the lower complexes of deposits. The fault was apparently formed after the deformation of both series and must be

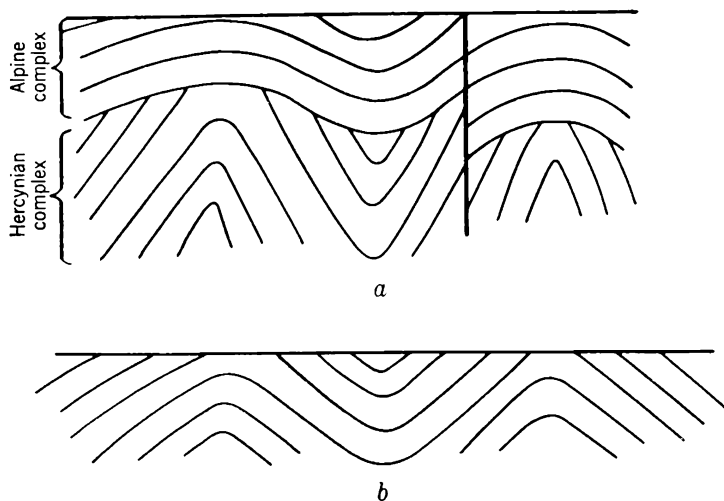


FIG. 60. Reconstitution of a former structure by subtracting the results of later deformations: (a) the present structure; (b) the structure that existed before deposition of the upper strata, restored by straightening the strata until the upper layers became horizontal.

“subtracted” when the upper stratum is restored to its original position. This “subtraction” will bring the displaced parts into the position they had before the faulting (Fig. 60b).

Figure 61 shows a more complicated case: a fault is localized within certain layers of the lower series. Since faults always are local disturbances dying out in all directions, this fault might have been formed at any stage in the lower series, either before the deposition of the upper or afterward. The answer to this question depends on a knowledge of the regularities governing the combinations of various structural forms. These regularities may show convincingly that the fault is physically associated with folds formed during the earlier tectonic stage, which

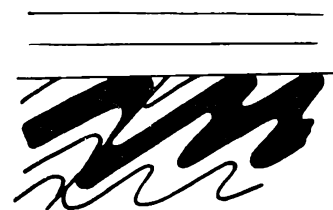


FIG. 61. An angular unconformity and a fault formed before the deposition of the upper strata.

deformed the lower series before the deposition of the upper. This fixes the time at which the fault was formed.

Similar questions arise in studying intrusive bodies. Figure 62a, for instance, shows an intrusive, the eroded top surface of which is uncon-

formably overlain by the upper series. It is clear that the intrusion is older than the upper stratum and does not have any relation to the structural forms in the latter. On the other hand, we must find its relations to the structure of the lower stratum. Figure 62*b* shows an intrusive that has penetrated both the upper series, and its origin is related to structural forms of the upper stratum. In Fig. 62*c*, an intrusive body occurs within the lower series and does not reach the contact between the series. Here the dating of the intrusive may be based on indirect evidence, such as contact metamorphism in surrounding rocks, which may involve the upper series and thus prove that the intrusion took place after the latter's deposition, or on the data from adjacent areas, where the relationship between the intrusive and the deformation of the structurally different series may be much clearer, or, finally, upon a knowledge of the laws governing the distribution and formation of intrusives in certain kinds of tectonic deformations.

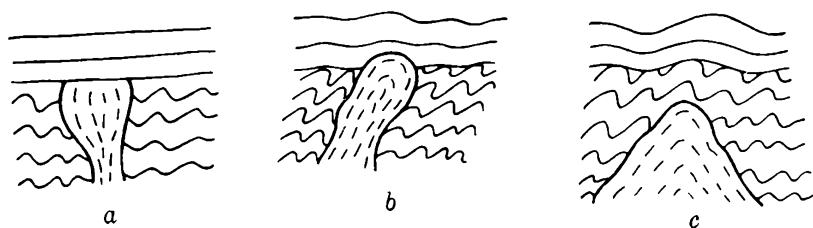


FIG. 62. Angular unconformities with different relationships to intrusives: (*a*) intrusive formed before deposition of the upper series; (*b*) intrusive formed after deposition of the upper series; (*c*) intrusive body within the lower series.

This section has described certain aspects of historical combinations of various structures, since the occurrence of structures in rocks may be correctly understood only if one knows whether or not the structures in question were formed simultaneously or at different times. The discussion below of mechanical combinations of structures will consider structures developed simultaneously and disregard the complications caused by superimposed tectonic stages.

MECHANICAL COMBINATIONS OF STRUCTURES

Tectonic forces affecting one part of the earth's crust or another form simple structures of the same type and approximately equal size only in exceptional cases. In the great majority of cases, the tectonic forces in the earth's crust produce combinations of various structural forms that complicate each other. In these combinations both deformations of a single type (such as folding) and deformations of different types (folding and faulting) are to be found. Superimposed and interrelated

dislocations of various kinds and sizes may be encountered in any part of the earth's crust.

Figure 63 shows a number of folds reflecting dislocations of a single order. But within each fold the individual beds reveal dislocations of other orders as well: they are broken into lenses, thinned out on the flanks, and thickened at the crests and troughs of the folds.

Figure 64 shows a bend (flexure) through a series of beds that are additionally disturbed in places by faults. The latter are clearly controlled by the overall bending and are much smaller than the flexure. A general description of the structures in a given part of the earth's crust

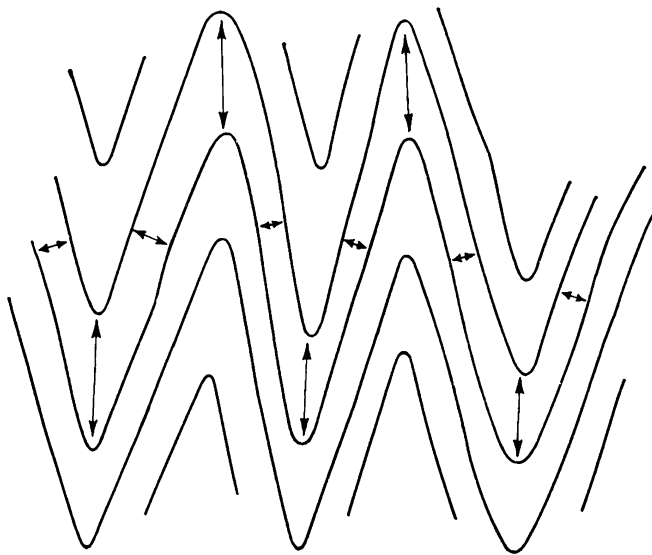


FIG. 63. Similar folds with regular increase in the thickness of the beds in the crests and troughs.

is adequate, if the folds in all the beds are characterized by a single type of bending, and does not require details. Upon making detailed studies, however, one must take account of the presence of the disruptive dislocations confined within particular beds and the complications caused by the superimposition of deformations of one order upon those of another. A detailed investigation, moreover, will reveal the presence of even smaller deformations, such as those reflected in the localized shrinkage of the thicknesses of certain beds in flexures and in the crushing of the rocks in fault planes. Thus when the study passes from dislocations of a general nature to particular deformations, it will have to take account of internal deformations within certain beds, with deformations that may, despite their small extent, be superimposed on each other, or, to put it more clearly, occur within each other.

A combination of dislocations of various orders can by no means be considered a coincidence. Such dislocations are closely and organically related and, on different scales, reflect a single process of deformation. In the structure shown in Fig. 64, the small faults are not just coincidentally developed in the flank of the flexure: they are the result of the flexure and reflect the peculiarities of the mechanism by which it was formed. In the same manner, the shrinkage and thickening of beds shown in Fig. 63 do not occur in folds coincidentally; their appearance is related to the mechanism of the folding. The external and internal dislocations in the rocks are thus related in accordance with regular laws.

The circumstances under which various combinations of dislocations arise will be described later. This discussion, restricting itself to the forms of these dislocations, will briefly review and classify the most typical combinations of different complex structures as reflected in various types of dislocations, beginning with the largest forms and passing to smaller ones. A survey will be made first of anticlinoria and synclinoria and then of anteklises and syneclises.

In regions whose structure is, in general, characterized by an alternation of anticlinoria and synclinoria, it is usually possible to distinguish anticlinoria and synclinoria of different orders superimposed on each other. For example, the main Caucasus range as a whole is an anticlinorium of the first order. Its generally anticlinal form is evident from the distribution in zones of rocks of different ages, the oldest at the core of the range and the younger ones on the flanks. This great anticlinorium, however, is complicated by a number of smaller ones separated by synclinoria. V. Ye. Khain (1949), for example, has distinguished seventeen anticlinoria and synclinoria of the second order within the Greater Caucasus (Fig. 65).

Many of these second-order anticlinoria and synclinoria have a trend that is oblique to that of the principal anticlinorium, and they are located en echelon relative to each other. This disposition of smaller anticlinoria and synclinoria is very common and apparently a typical phenomenon. Study of the Kura depression reveals that it is structurally a large synclinorium complicated by a number of structural forms of the next order: a structural uplift with outcrops of older rocks in its

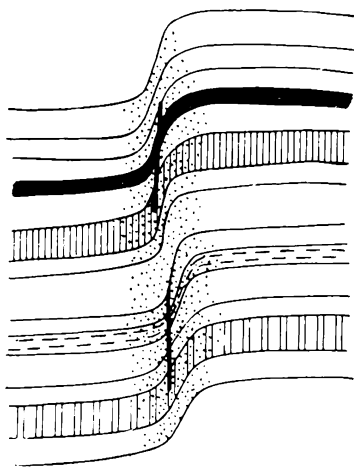


FIG. 64. A flexure that becomes a fault in certain beds. Dots indicate parts where the beds are compressed and fractured.

core trends along the entire length of the depression at a slight angle to the latter's axis, from the eastern end of the Trialetskirge Mountains at Tbilisi toward the Shemakha district, forming the so-called Chatminsk second-order anticlinorium. Thus the Kura first-order synclinorium is divided by the Chatminsk anticlinorium into two synclinoria of the second order, the Kartaly-Iora and the Palantyukyan synclinoria. Another anticlinorium of the first order, the Somkhet anticlinorium, is located farther south; here the en echelon disposition of the second-order anticlinoria within it is especially clear. Similar anticlinoria and

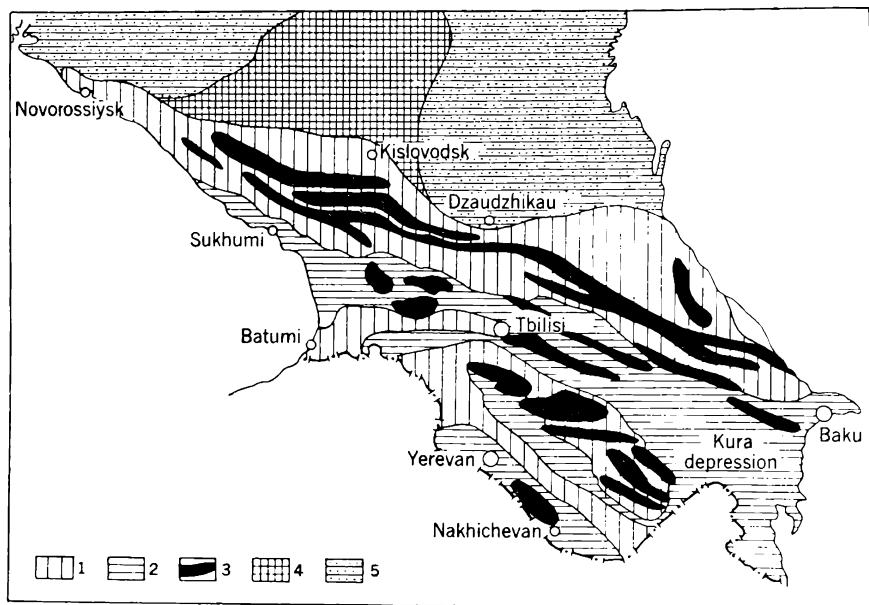


FIG. 65. Structural diagram of the Caucasus: (1) anticlinoria of the first order (meganticlinoria); (2) synclinoria of the first order (megasyntinoria); (3) anticlinoria of the second order; (4) antecises; (5) syneclises (after V. Ye. Khain).

synclinoria of different orders may be distinguished within any region where large structural forms are developed.

The anticlinoria and synclinoria of the first order are usually called *meganticlinoria* and *megasyntinoria*, reserving the terms "anticlinorium" and "synclinorium" for such structural forms of the second order. Anticlinoria and synclinoria are always complicated by folding, which differs between the two. For this reason they will be discussed separately.

Earlier in this book anticlinoria and synclinoria were divided into inverted and noninverted types. This division is reflected in the nature of the folding that complicates them. The most intensive folding occurs in inverted anticlinoria, a good example of which is the Greater

Caucasus range. The beds composing the anticlinoria predominantly form continuous folds. The positions and the types of the fold within the anticlinorium are controlled by certain regularities.

First, the most intensive continuous folding is usually confined to the most elevated parts of the anticlinorium. Toward the plunges and the flanks, the folding, as a general rule, becomes less intensive, and the continuous folds pass into intermediate and even discontinuous folds. Figure 66 is a diagram showing the distribution of the various types of folding in the Greater Caucasus range. Continuous folding is here confined to the central part of the meganticlinorium, whereas toward its southeastward and northwestward plunges and toward the adjacent

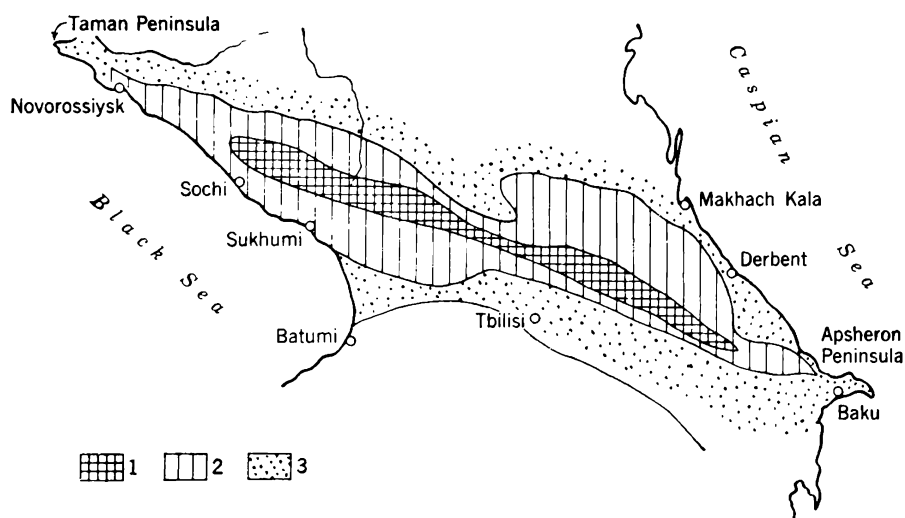


FIG. 66. Diagram showing the distribution of types of folding in the Greater Caucasus: (1) continuous folding; (2) intermediate folding; (3) discontinuous folding.

depressions one notices a change in the structural forms: the tightly crumpled folds with equally developed anticlines and synclines give way to intermediate folding with sharp anticlines and broad synclines, and these, in turn, pass into discontinuous folds consisting of separate domical uplifts. The discontinuous folds typically occur at the Apsheron and Taman peninsulas, which are the two opposite plunges of the Caucasus meganticlinorium.

Second, if an anticlinorium is asymmetrical, continuous folds are developed more intensively in its steeper flank, whereas on its more gently dipping flank the zone of intermediate folding begins very close to or even at the anticlinorium's axis. The southern flank of the Caucasus anticlinorium is much narrower; in this flank the continuous folding is most complicated and intensive. Here one encounters tight isoclinal folds, whereas the folds in the northern flank are more gentle.

Third, the continuous folds in the flanks of anticlinoria usually show a regular inclination: they are inclined and overturned in both flanks toward the adjacent depressions, and as a whole they resemble a fan (Fig. 67). If an anticlinorium is asymmetrical, the fan shape is also asymmetrical. For example, the folds in the southern flank of the Greater Caucasus range, which incline southward toward the Kura depression, are well pronounced along nearly the entire length of the range, whereas the folds in the northern flank are not always overturned to the north and, if they are, this characteristic is much weaker.

These observations are to a certain degree true both of anticlinoria of the first order (meganticlinoria) and of secondary anticlinoria. In different cases, therefore, the relationships between the structural forms of different orders may vary greatly; the regularities deriving from the larger structural forms are in places suppressed by those of the secondary structures, and in other places the reverse is true.

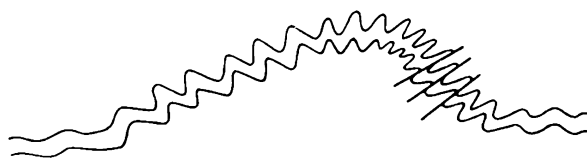


FIG. 67. Diagram of a fan-shaped anticlinorium.

The folds are usually parallel to the trend of the anticlinoria. If an anticlinorium of the first order is complicated by second-order anticlinoria whose trend differs somewhat from that of the former, the folds are usually parallel to the trend of the second-order structure and also en echelon with the meganticlinorium.

Anomalies are frequently observed in the plunges of both the principal and the secondary anticlinoria. In places the folds cross the trend of the anticlinoria, especially in the plunges, as in the Somkhet anticlinorium, the Kalba Range, the Pyrenees, and other places. These transverse folds are important in understanding the mechanism of continuous folding.

Noninverted anticlinoria are characterized by gentler folds complicating their structure. Here continuous folds occur rarely. Noninverted anticlinoria most commonly contain intermediate or discontinuous folds in the form of separate domical upwarps of the strata, such as those known in the noninverted anticlinoria of Miskhan in the Transcaucasus.

The continuous folds developed in the flanks of noninverted anticlinoria, as a rule, are overturned toward the axis of the anticlinorium and in combination form fold systems resembling an upside-down fan. Such, for example, is the structure of the noninverted anticlinorium of

the Ural-Tau, toward whose axis the folds of the Zilair synclinorium immediately to the west are inclined.

Synclinoria, both inverted and noninverted, unlike inverted anticlinoria, have less complicated fold structures. Here, as in noninverted anticlinoria, intermediate and discontinuous folds prevail. For example, the Kura megasyntinorium is characterized by brachyantclines and discontinuous domical uplifts, and similar folds are found in the Araksinskiy synclinorium in the Transcaucasus.

Beside folds, anticlinoria and synclinoria are complicated by faults. These are faults whose position and size are clearly related to the anticlinoria and synclinoria as a whole and are formed as a result of the same tectonic deformation that produced the large structural forms. Particular faults associated with individual folds will be reviewed later.

Large tectonic faults are, in the main, connected with noninverted anticlinoria and occur principally as reverse or normal faults. Their

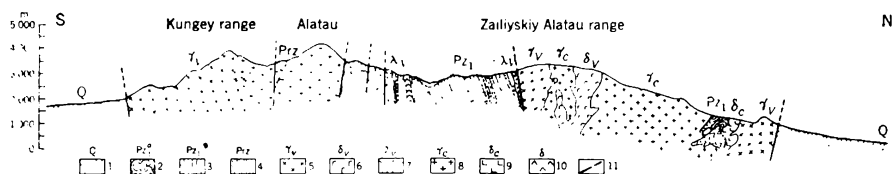


FIG. 68. Generalized section through the mountain ranges of the Zailiyskiy Alatau and Kungey Alatau in northern Tien Shan: (1) Quaternary sediments; (2) lower series of the Lower Paleozoic; (3) Lower Paleozoic; (4) Proterozoic; (5) Variscian granitoids; (6) Variscian gabbroids; (7) Variscian intrusive porphyrites; (8) Caledonian granitoids; (9) Caledonian gabbroids; (10) gabbroids of unknown age; (11) faults (after A. V. Goryachev and V. N. Krestnikov).

development depends on the amplitude of the uplift of the anticlinorium relative to the adjacent synclinoria and on the physical properties of the rocks involved.

Hard crystalline rocks and anticlinoria that are greatly elevated relative to the adjacent synclinoria are the most likely locations for large faults. Here large normal faults develop, usually parallel to the trend of the anticlinorium, turning its flanks into steps of blocks displaced vertically relative to each other. This structure is typical, for example, of the anticlinoria in the Tien Shan Mountains. Since they are greatly elevated over the adjacent depressions, the flanks of these anticlinoria are broken by numerous steplike reverse faults that have parallel strikes or intersect each other at acute angles (Fig. 68). Many of the reverse faults dip steeply toward the axis of the anticlinorium (common reverse faults), but there are also vertical reverse faults and those dipping in the opposite direction.

Faults may also disrupt the internal structure of noninverted anticlinoria. Sometimes anticlinoria are composed of a mosaic of blocks dis-

placed vertically relative to each other by numerous normal and reverse faults.

A large number of steep faults cut through the interior of the anticlinorium in the central Armenian upland. The Late Tertiary and Quaternary volcanic activity in this region is associated with the faults. The positions of these extinct volcanos frequently reveal the location and the strike of faults that are not exposed at the surface because of lava flows. For example, a north-south trending chain of Tertiary and Quaternary volcanos in the Abul-Samsarskaya Ridge near the town of Akhalkalaki indicates the presence of a large tectonic fissure covered by lava flows.

Numerous faults intersect the anticlinoria of the Tien Shan Mountains and in some places break them into blocks. The structures, in the northern Tien Shan, of the Zailiyskiy Alatau and the Kungey Alatau Ranges



FIG. 69. Diagram of harmonic (a) and disharmonic (b) folding.

(Fig. 68) are very interesting. Here a number of longitudinal normal faults cut off a long narrow strip along the axis of a large noninverted meganticlinorium and displace the strip vertically, forming the deep Kebino-Chilik graben. The interiors of synclinoria, according to available data, do not have such large faults.

Faults are also rare in inverted anticlinoria, but they may occur within or near large bodies of crystalline rocks. For example, a number of large longitudinal reverse faults are encountered in the Greater Caucasus range, especially in the northwestern part, where some areas are broken up by such faults into narrow plates elevated or lowered relative to each other.

A. V. Peyve has suggested the term *deep fractures* for tectonic faults of the first order whose development is closely related to that of the largest structural forms. Their significance will be discussed below.

We may now review the dislocations that complicate folded structures, beginning with continuous folding. Regions of intensive continuous folding abound in disharmonic folds. The disharmony is expressed in the fact that, in a series of strata folded during a single period of

diastrophism and by a single tectonic deformation, the structures differ at different levels and in different suites and beds. Introductory textbooks of structural geology distinguish between harmonic and disharmonic folds. The former are folds having approximately the same pattern in all the layers of the folded series (Fig. 63). If different groups of beds have different fold patterns—for example, if one of them contains additional smaller folds (Fig. 69)—the folding is disharmonic.

What is known of the structures of folded strata suggests that harmonic folding is a local phenomenon, always limited to small parts of the earth's crust. Harmonic folds can be found within a single outcrop, but if one considers the entire folded stratum as a whole, one realizes that disharmony is a typical phenomenon in every region of continuous folds. The disharmony may be manifested on different scales and divide the folded stratum into separate structural complexes of greatly different magnitude.

Disharmonic folds may frequently be confined to a particular group of beds and extend neither upward nor downward in the section. Figure 70

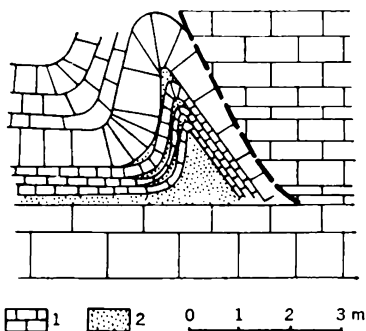


FIG. 70. Disharmonic folds in almost horizontal beds in the Rion River valley in Georgia: (1) limestone; (2) marl (after I. V. Kirillova).

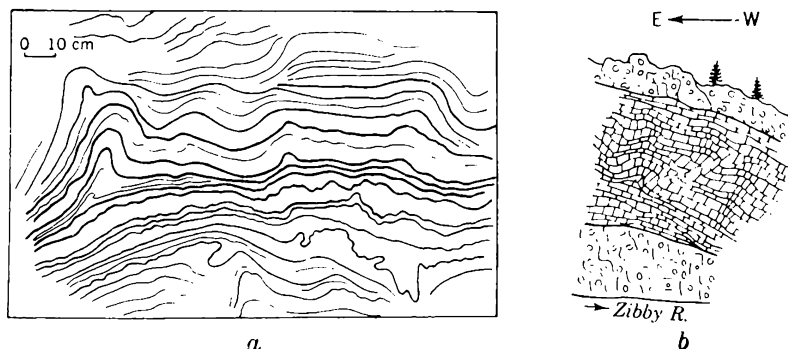


FIG. 71. Examples of complex disharmony: (a) vertically plunging small disharmonic folds in gneisses of Karelia (after A. A. Sorsky); (b) disharmonic folds in platy carbonaceous limestones in the Zidda Valley of Tadzhikistan (after I. V. Kirillova).

shows such a fold, which is isolated within a monoclinaly folded stratum. It is easy to see that the fold was formed through distortion in certain beds, which became thinner in some places and thicker in others.

Figure 71 shows a folded structure with complicated disharmony: different beds or groups of beds are folded to varying degrees, and

this again is related to a very complicated distortion of the beds, which frequently change in thickness.

The degree of disharmony may vary widely. It depends both on the intensity of the folding and on the physical properties of the rocks involved. Plastic and thin-bedded rocks, such as clays, shales, thinly laminated marls, and limestones, reveal greater disharmony and become complicated by a greater number of additional small folds than strong and massively bedded layers, such as sandstones and massive limestones.

There are abundant cases in which folds of various magnitudes superimposed on each other follow the same trend. The trends, however, are not always parallel. Frequently there are more complicated combinations of folds with different trends. N. P. Semenenko (1946), for example, has described small disharmonic folds complicating the form of large Precambrian folds and striking across or diagonally to the latter's trend. The additional cross folds and diagonal folds greatly affect the concentration of iron mineralization. M. V. Gzovsky, for example, has observed cross folds of secondary orders, developed in large folds of Devonian deposits in the Kara-Tau Mountains. Disharmonic cross folds in Archean sediments are very abundant, for instance, in Karelia.

Cross folds are especially abundant in the vertical or very steeply dipping flanks of large folds. In such cases, the additional cross folds accordingly plunge vertically or steeply.

Some of the above examples indicate that continuous folding is usually accompanied by distortion of the beds, especially in the form of local changes in thickness. Even in places of harmonic folding the beds become distorted: they become thicker in the crest and troughs of the folds and thinner on the flanks. Folds in which no distortion of the beds appears (so-called concentric folds) are very rare; they may occur, for example, as small folds bent very gently. Continuous folds, as a general rule, are "similar": all the folds have the same radius of bending. This is possible only if the beds change in thickness as described above (Fig. 63).

If folds of various magnitudes are superimposed disharmonically, changes in thickness and extremely complicated distortion of beds may be seen in folds of all sizes.

Continuous folds are usually broken by faults whose origin is related to that of the folding. The continuous folds are most commonly combined with overthrusts and, to less extent, with strike-slip faults. Thrusts are largely confined to overturned and recumbent limbs whose dip corresponds to that of the axial planes of the respective folds (Fig. 72). Thrusts may be associated with folds of various orders and accordingly have widely varying amplitudes of displacement from a few centimeters to ten or fifteen kilometers. There are also low-angle thrusts

that cut not one but a group of folds and displace the hanging wall, producing a plate of cutoff folds separated from their continuation in the foot wall. Such a thrust or overthrust is called an *intersecting thrust* (Fig. 72). The fault planes dip steeply in some places and gently in others. If a number of overturned folds have parallel trends, so-called imbricate thrust faults may be formed (Fig. 72).

Some years ago extensively developed overthrusts of great extent and displacement, reaching tens or even hundreds of kilometers, were believed to exist in the Alps and certain other mountainous regions. Some

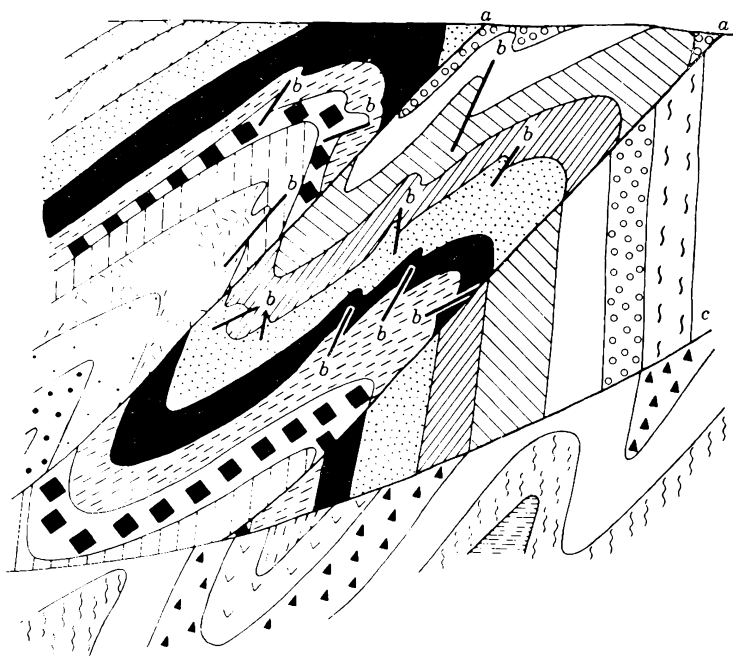


FIG. 72. Overthrusts: (a and b) thrusts associated with folds of various orders; (c) intersecting overthrusts.

geologists believed also that nappes occurred frequently in Soviet mountainous regions such as in the Transbaikal, the Urals, the Kara-Tau, and others. The latest studies have disclosed, however, that no such tremendous horizontal thrust sheets exist in the U.S.S.R.; the conceptions based on them are the result of methodologically erroneous interpretations of the geologic data. Moreover, recent investigations show that the conceptions of tremendous nappes in the Alps are exaggerated and that the structure of the Alps should be interpreted much more simply (Beloussov, Gzovsky, and Goryachev, 1951). The greatest displacement appears in thrusts (charriage) of gravitational origin (see below).

The strike of overthrusts is nearly everywhere parallel to that of the folds with which they are associated. Local deviations occur, but they

are not important. Since each overthrust is confined to a certain fold or group of folds, it does not extend further than the respective folds or group of folds and dies out where the latter die out.

Strike-slip faults, whose development is connected with continuous and intermediate folding, usually have strikes diagonal to the fold's trend (Fig. 73). Transverse strike-slip faults also occur. Large strike-slip

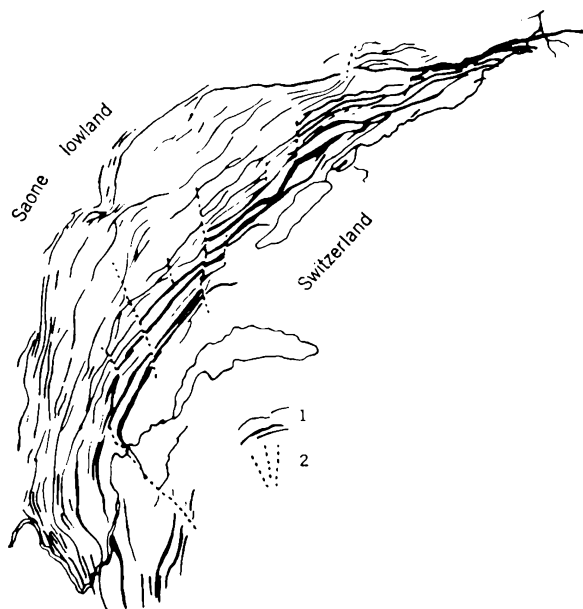


FIG. 73. Strike-slip faults in the Jura Mountains: (1) axes of anticlines; (2) strike-slip faults (after A. Heim).

faults are relatively rare, apparently, but among dislocations of less size they may be abundant.

One internal dislocation regularly associated with continuous folding is flow cleavage. Cleavage develops under conditions of intensive folding; the types of cleavage and their disposition in relation to folds were discussed in the preceding chapter.

Anticlinoria are complicated not only by folds but also by faults and frequently by igneous intrusions. The latter are especially abundant in inverted anticlinoria, and above all in regions of continuous folding. Here large granite batholiths are typical of the most uplifted parts of inverted anticlinoria. Sill-like basic intrusives are widely developed in the lower complexes of deposits. All the rocks may be cut by fissure intrusions, and volcanos of the central type appear with great violence on the earth's surface.

The flanks of inverted anticlinoria that contain folds of the intermediate type are typical localities for "minor" intrusives, especially for those of alkaline composition. Other forms of igneous activity are much

less frequently encountered here. Batholiths are not typical of noninverted anticlinoria. Here the intrusives are mainly of the fissure type, and they sometimes take the form of "minor" injections. Surface volcanism is very intensive.

Synclinoria almost never contain manifestations of igneous activity; the rare igneous bodies are extrusives and "minor" intrusives.

We may now turn to anteklises and syneklises, which, like anticlinoria and synclinoria, may be of various orders. For example, the Canadian shield, which is a great anteklise, is complicated in its center by a syneklise of the second order, in which Hudson Bay is located. The Voronezh massif, which is also an anteklise, contains many second-order upwarps and depressions.

The flanks of anteklises and syneklises are frequently complicated by flexures and by steeply dipping local benches and normal faults, which give the flanks the character of ladders. In this respect the structure of the Dnepr-Donets Basin, which is a typical syneklise, is most interesting: a number of normal faults have been found in the flanks of this basin, descending by steps toward its axis.

At one time A. P. Karpinskiy thought that all exposures of the crystalline basement of the Russian plain (an anteklise, according to modern terminology) were horsts bordered by normal faults (1883, 1919). Later, however, this concept was replaced by another—that the flanks of this anteklise dipped continuously without faults. The latest studies have shown, however, that, although some of the faults pointed out by A. P. Karpinskiy do not exist, combinations of faults, each of small amplitude, are very common in the flanks of the anteklise.

Some anteklises are broken up by closely spaced normal and reverse faults. One example is the central plateau of France, which is a mosaic of horsts and grabens formed by vertical faults. The largest dislocations in the nature of normal faults are associated with anteklises: these are the rift valleys of East Africa and the Upper and Lower Rhine grabens in Western Europe. The conditions under which these were formed will be discussed in Chap. 27.

Syneklises are typically complicated by discontinuous folds; in this respect they resemble synclinoria. In syneklises, however, the discontinuous folds in the majority of cases are more gentle than those in synclinoria. Nevertheless, very large discontinuous folds may occur in certain syneklises (such as the Zhiguli and Don-Medveditsa arches and the Rocky Mountains).

The discontinuous folds in syneklises, like those in synclinoria and in the flanks of anticlinoria, are characterized to the same degree by dislocations of various orders very much like those that occur in the larger uplifts, such as anteklises and even anticlinoria.

Large discontinuous upwarps are characteristically complicated by folds of the same discontinuous type but of smaller size. The discussion

of the general features of discontinuous folding has already indicated the superimposition of small domes on larger discontinuous folds.

Discontinuous folds are frequently complicated not only by smaller folds of the same type but also by folds that may be called *continuous* or *intermediate*. The latter occur in the form of elongated linear folds on the flanks of discontinuous domes, either as single or, more frequently, as groups of parallel folds resembling miniature fold zones. These folds are always very small: the largest are a few score meters wide but the majority no more than a few meters. On the other hand, they can be traced for several hundreds of meters or even several kilometers along their strike.

The extensive distribution of small linear folds complicating discontinuous folds was established by V. V. Bronguleyev (1951) through his special studies of the Russian platform. He also definitely established the tectonic origin of these folds, which had previously been considered exogenic (caused by glaciers, landslides, etc.). It was also found that the small linear folds within discontinuous folds are not disposed at random but show a tendency to form concentric belts around uplifts and are predominantly parallel to the structural contours (Fig. 74). Vertical sections show that the small linear folds are sharply disharmonic and are usually limited to only a few beds. Study of these folds, as will be seen later, provides valuable data on the mechanism of folding.

Disharmony in discontinuous folds appears in the formation of so-called diapir domes. Such diapir cores have been described earlier in this book as extreme local distortions in the thickness of beds, especially plastic beds. A diapir dome is a complex structural form, the core of which is the most important part. A diapir dome contains three structural complexes. A plastic complex (salt, gypsum, or clay) lies at the center, and its deformation takes the form of a very sharp local swelling of the bed. The upper complex is formed of the rocks overlying the plastic core. The roof becomes elevated above the swelling and usually acquires a domical structure; it may be partially or completely pierced by the diapir core.

The structure of the lower complex underlying the plastic beds is usually little known because of its great depth. The lower complex has been revealed by drill holes in areas of diapir domes in Bashkiriya. It was found that the shape of the surface of the lower complex, consisting of limestones of Artinskian and Upper Coal Measures age, shows no correspondence to the structure of the middle and upper complexes. Here the diapir cores are most often situated above depressions in the top surface of the lower complex and only rarely over elevations. Geophysical study in the Emba district, where diapir domes are very abundant, shows that the lower complex is entirely undisturbed, apart from a general monoclinal dip toward the Caspian Sea. Thus, dia-

pirs are sharply disharmonic structures originating in a subsurface flow of the material from some parts of plastic beds toward others.

The most gentle discontinuous uplifts are usually covered and not disturbed by faults, but the largest upwarps of this type are frequently

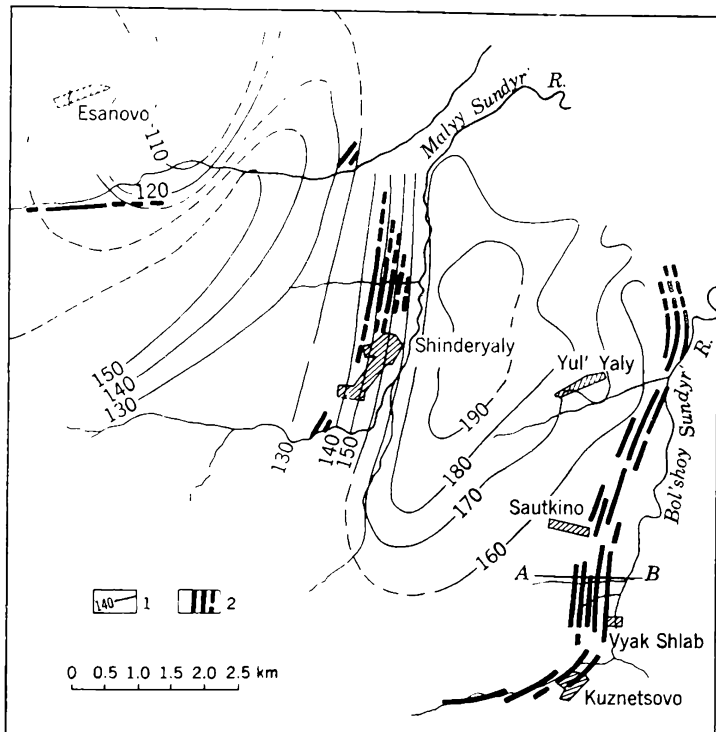


FIG. 74. Small secondary folds in the eastern part of the Russian plain. A district in the basin of the Sundyr' River (structural diagram): (1) contour lines on the top of one of the marked strata in the Tatarian suite; (2) small folds.

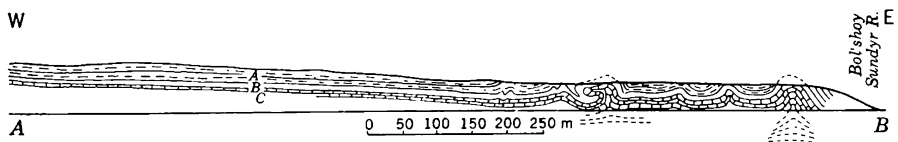


FIG. 75. Small secondary folds in the eastern part of the Russian plain. A district in the basin of the Sundyr' River (after V. V. Bronguleyev). Cross section along A-B (see Fig. 74). The various strata of the Tatarian suite are shown separately.

associated with both normal and reverse faults. The faults break up some upwarps to such an extent that they are called *broken plates*. Several examples of domes broken by normal and reverse faults have been mentioned earlier. If the flanks of the discontinuous fold are steep

(as in a box fold), the upthrusts associated with these flanks are the ultimate development of flexures. In other cases the entire dome is broken up by faults that fall into two systems: longitudinal and radial. Some regularities in the distribution of faults on uplifts will be discussed later.

Igneous activity in areas of synclises, anteklises, and discontinuous folding is generally much less extensive and, above all, less diverse than in anticlinoria. Many structural uplifts and basins contain no syntectonic igneous bodies. Examples of such structures are the Moscow basin and the Voronezh anteklise. In other cases, however, there are manifestations of igneous activity. In synclises this most frequently takes the form of sills and basic lava flows, which are sometimes of large extent, and of fissure and minor intrusions. Flows originate in fissures of regional extent and may cover tremendous areas. The fissure and minor intrusions are usually associated with discontinuous folds and especially with faults subordinate to these folds. Extensive surface volcanic activity is frequently associated with anteklises.

The Tunguska basin, with its numerous sill-like, fissure, and "minor" intrusions and blanket flows, is an example of a synclise whose formation was accompanied by violent igneous activity (during the Late Paleozoic). Enormous extrusions of basaltic rocks took place during the Mesozoic in the synclises of India (the Deccan traps) and in South America (the Paraná depression). There was also intensive surface volcanic activity at the end of the Tertiary and beginning of the Quaternary in the anteklises of the central plateau of France, in the Rhine mountains, and elsewhere. The relationship of "minor" intrusions and fissure intrusions to large discontinuous folds is especially clear in the Rocky Mountains, which were folded and intruded during the Tertiary period.

REGIONAL COMBINATIONS OF STRUCTURES

Observations have shown that the various structures in rocks are not randomly distributed in the earth's crust but follow a certain regularity. These regularities, however, emerge only if, in addition to the morphology, one considers the age of the structure, meaning the structural stage with which they are associated. For example, one may compare the extremely complexly dislocated Precambrian rocks of the Baltic crystalline shield with the relatively undisturbed Paleozoic and younger deposits in the center of the Russian plain and distinguish the areas in which these rocks appear as regions in which different structural conditions prevailed. Such a comparison, however, does not present a true picture of the regional distribution of structures formed at the same time: it will reflect not the regional distribution but the historical suc-

cession of their formation, since the Precambrian rocks of the Baltic shield and the younger deposits of the Russian plain belong to structural stages of different ages and are separated by a sharp angular unconformity. On the other hand, the structure of the complex of Paleozoic deposits on the Russian platform is completely comparable with that of the rocks of the same age in the Urals, whose structure was formed simultaneously with that of the Russian platform. Thus one must here distinguish the regularities in the distribution of different structures formed at the same time.

It will become clear later that, although the tectonic process, which encompasses different types of movement, proceeds continuously and thus continuously alters and develops the structure of the earth's crust, certain stages in the latter's history may be distinguished as stages of great structural transformation, in which large numbers of new, sharply distinct structures were produced. These periods are characterized by especially intensive folding developed in certain zones. Thus they are usually called *periods of folding*. Other processes in the earth's crust, however, such as vertical movements, faulting, and igneous activity, are also intensified during these periods.

There were many periods of folding in the Precambrian. Of these, only the latest two or three are known, since the traces of earlier ones have been obliterated by Precambrian metamorphism. Still later periods of folding, from the beginning of the Paleozoic down to the present, are well known. During this time, lasting about five hundred million years, there were three principal periods of large-scale structural transformation. One of them took place at the end of the Early Paleozoic, mainly in the Late Silurian, the second in Late Paleozoic (the Carboniferous and Permian periods), and the third at the end of the Mesozoic and mainly during the Tertiary. These stages of folding have been named, respectively, the Caledonian, Hercynian (or Variscian), and Alpine.

The segment of geologic history from the end of one period of folding to the end of the next one is referred to as a *geotectonic stage* or *geotectonic cycle*. These cycles have the same names as the respective periods of folding. Thus, beside the Precambrian cycles, which will not be discussed here, there are the Caledonian cycle, which encompasses the Early Paleozoic (Cambrian and Silurian period), the Hercynian or Variscian cycle (Devonian, Carboniferous, and Permian), and the Alpine cycle (Mesozoic and Cenozoic).

The significance and content of the geotectonic cycles will be mentioned later. Here it must be pointed out that to understand the regional distribution of structures, one must compare these geotectonic cycles and the structures of the rocks as they were produced during a given cycle. Thus the structures or combinations of structures formed during the Caledonian cycle may be called Caledonian structures, the

structures or combinations of structures produced during the Hercynian cycle may be called Hercynian, and those formed during the Alpine cycle may be called Alpine.

The distinction between Caledonian, Hercynian, and Alpine structures requires certain methodologic rules. Since the Alpine geotectonic cycle is the latest in geologic history, the structural forms that arose in this cycle have remained preserved in their pure form without distortion and alteration by later movements in the earth's crust. They may be observed directly. But they can be seen only in a complex of rocks formed during the Alpine cycle, that is, in Mesozoic and Cenozoic rocks, since only the structure of these rocks will give a true picture of the disposition of Alpine structures. One cannot see the Alpine structures in their pure form by studying older strata, Paleozoic and Prepaleozoic, since these might have been deformed during preceding Hercynian, Caledonian, or even Precambrian cycles. For example, conclusions in regard to the Alpine structures in the Russian platform within the Donets basin must be based upon the structures in the Jurassic, Cretaceous, and Tertiary sediments, but not upon those of the Carboniferous, for the Paleozoic sediments had already been deformed during the Hercynian cycle and their folds are not directly related to the Alpine structural forms. As can be seen in the Mesozoic and Cenozoic strata, the Alpine structures are here very gentle, taking the form of a number of gentle domical uplifts differing substantially from the intense folding of the Paleozoic.

Only in those cases where Mesozoic and Cenozoic strata conformably overlie the Paleozoic complex and the latter was deformed together with the former during the Alpine diastrophism may one, to a limited extent, use the structures of the Paleozoic strata in studying Alpine structures. But interpretations and conclusions based upon such analogies cannot be considered proofs.

Determination of the structures formed during older cycles is more difficult, since these structures became deformed and altered during later cycles as well. Hercynian structures were redeformed and altered by the Alpine movements and the Caledonian structures by the Hercynian and Alpine movements. Consequently, a study of the present structures in the Hercynian complexes (Devonian, Carboniferous, and Permian) is not adequate for a reconstruction of the original Hercynian structures: from the present structures we must first "subtract" the results of the later Alpine tectonic movements. The Caledonian structures may be reconstructed by "subtracting" the results of the Alpine and Hercynian structural alterations superimposed on the Caledonian complex (Cambrian and Silurian). It has already been shown how this "subtraction" should be carried out: by straightening both series until the younger strata become horizontal and the older, being straightened to

the same extent, are restored to the position they had before the later deformation took place.

Experience has shown that in regions where the younger strata underwent only gentle upwarping and downwarping that produced antecises and synclises or gentle discontinuous folds while the underlying strata were intensively deformed by older tectonic movements that produced sharp continuous folds, there is no need to "subtract" the younger deformations. If the intensity of the latter is comparable to that of the preceding tectonic cycle, however, the subtraction must be made.

A study of the regional distribution of Caledonian, Hercynian, and Alpine structures shows that the earth's crust has two principal types of structurally different regions: *folded zones* and *platforms*. The two types can most easily be recognized in the Alpine complexes. In some regions the Mesozoic and Cenozoic deposits are intensively deformed in certain areas, which are predominantly elongated belts or zones, where they form great uplifts and downwarps—anticlinoria and synclinoria—and tightly compressed folds that are primarily of the continuous type. In these regions the deposits are penetrated by numerous intrusives, among which batholiths occupy a prominent position.

In other regions, the sedimentary deposits are much less disturbed, forming very gentle antecises and synclises, which are complicated to a small degree by faults and mainly by gentle discontinuous folds; igneous activity, if any, is relatively weak.

Examples of the two types of regions are the Caucasus Range, with its intensively deformed Mesozoic and Cenozoic strata, on the one hand, and the Russian platform, where the same deposits are nearly horizontal except for very gentle antecises and synclises and individual discontinuous folds. The first case is an example of an Alpine folded zone and the second of an Alpine platform.

The same distinction can be made in the case of the Hercynian and Caledonian cycles. For example, the deposits of the Hercynian complex (Middle and Upper Paleozoic) are intensively dislocated in the Urals, whereas on the Russian plain and in central Siberia their structures are very gentle. In this case, one must evidently speak of a Hercynian folded zone in the Urals and Hercynian platforms in the Russian plain and in Siberia. The lower Paleozoic in Norway forms a typical folded zone, with anticlinoria and synclinoria, continuous folding, and intensive igneous intrusions, whereas deposits of the same age in Estonia and Latvia are distinguished by their nearly horizontal occurrence, typical of platform conditions.

The Hercynian folded zones are best observed where platform conditions obtained during the Alpine cycle, so that the Hercynian structures were almost unchanged during the succeeding diastrophism. Similarly, the structures of the Caledonian folded zones are best seen in regions

that were platforms during the Hercynian and Alpine cycles. Wherever older folded zones, on the other hand, have been covered by younger zones, the older structures are very strongly altered.

From this standpoint the Urals are an advantageous locality for studying the structure of the Hercynian folded zone, since in the Alpine cycle this area was part of a platform, where the Mesozoic and Cenozoic deposits remained almost in their primary attitudes. The Scandinavian mountains, likewise, are suitable for studying the structure of the Caledonian folded zone, since neither the Hercynian nor the Alpine cycles produced any intensive dislocations here.

Younger folded zones usually occupy smaller areas than older ones. Consequently, the older folded zones are partially incorporated into younger platforms; in these parts the structures of the older folded zones remain fairly well preserved. Older folded zones, however, also exist in regions in which intensive younger dislocations have later taken place. Here the older structures, being intensively reformed, can be reconstructed only partially and with great difficulty.

The large structures of folded zones are anticlinoria and synclinoria. These are complicated by folding that follows the regularities mentioned above. The regular features are the development of continuous folding in the center of inverted anticlinoria, a fanlike structure, and a transition from continuous to intermediate and discontinuous folding toward the outer edges of the anticlinoria and the adjacent synclinoria. Another regular feature is the more gentle folding (intermediate and discontinuous) of synclinoria and noninverted anticlinoria, the latter being accompanied by "deep faults" and having a reverse fanlike structure. Folding of all types is regularly complicated by cohesive and disruptive dislocations of successive orders superimposed on each other, including dislocations in the internal texture of the rocks.

Inverted anticlinoria usually contain large batholiths, as well as fissure intrusions, whereas noninverted anticlinoria and synclinoria are areas in which minor intrusions and fissure intrusions predominate. Extrusive volcanic activity is especially intensive in noninverted anticlinoria.

The largest structures on platforms are anteklises and syneklises. The anteklises are frequently dislocated by "deep faults," whereas the syneklises are regularly complicated by discontinuous folds, which, in turn, are complicated by cohesive and disjunctive dislocations of lesser orders. Igneous activity on platforms is generally not intensive, but in certain cases the platforms are covered by thick lava flows and (in association with large discontinuous folds) contain minor and fissure intrusions.

Although folded zones and platforms differ sharply with respect to their typical structural forms, these two basic types of structural complexes are almost always regionally connected by certain fairly extensive transitional zones of intermediate structure. These zones usually coin-

cide with *foredeeps*, as the outermost synclinoria of folded zones, which border typical platforms, are called. The transitional zones are characterized by intermediate and discontinuous folding and by abundant diapir domes associated with thick deposits of salt and gypsum in the foredeeps. Igneous activity is usually weak and produces minor intrusions of alkaline composition.

Examples of Alpine transitional zones are the Terek-Kuban basin in the northern Caucasus, with its intermediate and discontinuous folding in the Grozny and Kuban oil-bearing districts and minor intrusions in the Mineralnyye Vody district. An example of a Hercynian transitional zone is the area west of the Urals, where intermediate and discontinuous folds, including diapir domes, are extensively developed.

There is also a gradual transition from a folded zone to a platform in the western foothills of the Appalachians (Fig. 76). A number of

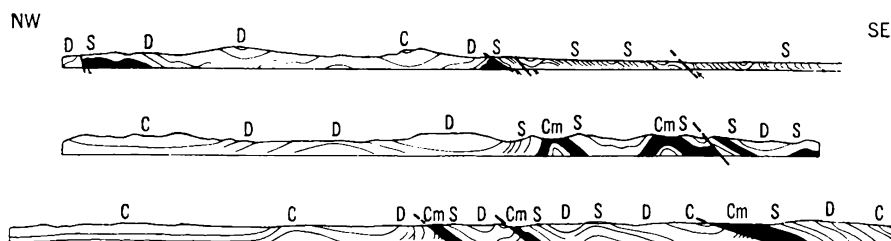


FIG. 76. Sections through the western slopes of the southern Appalachians. Transition from a folded zone into a platform. The central section is the continuation of the upper one but placed en echelon at the northwest.

sections show that folding with an equal development of anticlines and synclines gradually passes westward into comb-shaped folds with sharp anticlines and broad gentle synclines, and still farther west it becomes a typical platform with horizontal or slightly undulating beds.

Transitional zones are of great practical interest because of their concentrations of certain mineral deposits, such as coal and oil.

Since the area of the folded zones decreases from older geotectonic cycles to younger ones, and that of the platforms accordingly increases, the Alpine platforms are larger than Hercynian and the Hercynian platforms are larger than the Caledonian. As far as Precambrian rocks are concerned, these are everywhere intensively folded and penetrated by intrusions; if there were any platforms in Precambrian times, they were probably very small. Because of these peculiarities in the disposition of folded structures of various ages, below the nearly horizontal deposits of platforms there are always intensively dislocated complexes of older rocks representing folded zones of earlier cycles. There may be several variations. The Mesozoic and Cenozoic platform complexes may be underlain by Hercynian folded zones, in other words, an Alpine plat-

form rests on a Hercynian folded basement. Examples of this are the Urals, central Kazakhstan, and the Donets basin. In other cases, an Alpine platform complex may rest on a nearly horizontal complex of a Hercynian platform, which is, in turn, underlain by an intensively folded lower Paleozoic stratum, a folded zone of the Caledonian cycle. Thus one would have an Alpine platform on a folded Caledonian basement. Platform conditions existed, in this case, not only during the Alpine cycle but also during the Hercynian. Finally, below the Mesozoic and Cenozoic strata of the Alpine platform, the entire Paleozoic complex may lie horizontally and only the Precambrian may be intensively folded and metamorphosed. This would constitute an Alpine platform on a Precambrian crystalline basement. Platform conditions existed here as early as the Caledonian cycle and continued through the Alpine cycle. An example of this is the Russian platform.

Thus a platform is always characterized by two sharply differing structural stages: an upper, or platform, stage and a lower, or folded, stage. But the age of the folded basement may vary widely.

The map at the end of this book shows the positions of the principal structural elements in the earth's crust. First of all, the Alpine folded zones and Alpine platforms are distinguished. Next one may see the parts of the Alpine platforms that rest on folded basements of varying age (Hercynian, Caledonian, or Precambrian). Finally, the map shows the major anteclines and synclines and certain other very large dislocations. This map will be discussed in greater detail later.

THE PECULIAR NATURE OF DISLOCATIONS IN ARCHEAN STRATA

The distinctions between continuous and discontinuous folds and between anticlinoria and synclinoria, on the one hand, and anteclines and synclines, on the other, can be seen in rocks of Cenozoic, Mesozoic, Paleozoic, and Proterozoic age. In regard to Archean, or the oldest, rocks, however, these terms are hard to apply. Archean rocks predominantly contain dislocations of peculiar forms that may be studied in the crystalline shields, such as the Baltic shield and the Canadian shield, where the Archean rocks were dislocated only during Archean times and remained nearly undisturbed by later tectonic movements.

The Archean gneiss and schist series form very large, typically arch-like structural uplifts whose shapes are circular, oval, or irregular in plan. These uplifts are closely associated with the intrusion of granite batholiths, which are usually encountered at the core. The Archean uplifts are as much as several scores or even hundreds of kilometers in diameter. The relative elevation of their strata is not known, but it must have been very great (on the order of several kilometers). The gneisses usually dip steeply, often vertically, on the flanks or are even overturned.

If the flanks are overturned, the uplifts become mushroom-shaped. In the oval-shaped uplifts, both the long and short flanks have been found overturned. These overturned Archean uplifts differ substantially from the discontinuous folds known in younger strata.

The gneiss domes are usually intensively complicated by small additional folds showing sharp and complex disharmony (see Fig. 71). These additional folds are very small compared with the principal domes: they take the form of bendings ranging from several centimeters to several score meters in size. In general, these folds are continuous. Their hinges are usually vertical, so that their cross sections may consequently be seen in the horizontal plane, which reveals the bending of the vertical strata in the flanks of the large domes. Disharmonic folds with horizontal hinges are rare.

Besides the additional small-scale disharmonic folds, the Archean rocks contain abundant distortions in the form of shrinking and swelling of the beds and the formation of lenses. These internal dislocations are intensive. Faults formed in Archean times are small and local and are subordinated to the disharmonic folds, cutting through a few beds in the form of strike-slip faults. The crystalline shields may also contain large tectonic ruptures such as normal and reverse faults. These, however, being much younger, are not associated with Archean tectonic structures. The Baltic shield, for example, is broken up by fractures and normal faults that are largely of Tertiary age.

In their structure the Archean uplifts resemble diapir domes, but on an enormous scale and with cores composed of granite instead of salt or clay. The Swiss geologist Wegmann (1935) called these domes *granite diapirs*, extending the term "diapir" to all structural forms produced by the elevation or piercing of some rocks by others from below.

CHAPTER 10

TYPES OF TECTONIC MOVEMENT

The structures described in this section result from the displacement of the material in the earth's crust. Displacements of crustal material resulting in changes in the structure of the crust are called *tectonic movements*.

The variety of structures indicates that tectonic movements are also varied. Some result in large uplifts and downwarps, others cause crumpling of strata into small folds, and still others cause fracturing of the crust. It is very important, therefore, to devise a classification of tectonic movements that may be used in describing the tectonic history of the earth's crust. The construction of such a classification, however, involves great difficulties. They arise mainly from our still very inadequate knowledge of the origin of tectonic movements and of their primary causes. For this reason, it is impossible in the present state of geotectonics to speak of a genetic classification of tectonic movements.

The division of tectonic movements into radial and tangential, or essentially vertical and essentially horizontal, is widespread in geologic literature. Vertical uplift and subsidence of the crust and movements along normal and reverse faults may be referred to the first group, folding and overthrusting to the second. This classification, which at first glance looks quite rational, is too schematic and, apart from use in elementary teaching, is almost completely unsuitable. In the first place, it is incorrect to think of the principal displacements of crustal material as being vertical and horizontal. These displacements occur in any direction; the vertical and horizontal directions appear to be the main ones simply because we are in the habit of depicting the structures of the earth's crust either in vertical sections or on horizontal maps. In the second place, tectonic movements are never observed in pure form: they are always complex combinations of a variety of movements. Vertical movement always produces subsidiary horizontal and oblique displacements induced by the single vertical mechanism. This makes the problem of dividing movements into vertical and horizontal a less simple one than would appear at first, for it is difficult to determine what particular kind of movement is under consideration.

Difficulties in working out a classification of tectonic movements arise also from the fact that the individual types of structures, and therefore of the tectonic movements causing them, are not of equal significance. For example, faults are in many cases not independent but often result from either uplift or downwarping of the crust or from folding. It may even be stated that folding itself is a secondary phenomenon developing under certain conditions as a mechanical result of vertical movements of the crust. From this standpoint, it would be possible to divide tectonic movements into primary and secondary. The former would include upwarping and downwarping of the crust, the latter folding and tectonic fracturing. There is some basis for such subdivision, but at present it is largely an interpretation rather than a classification of observed facts. The terms *orogenesis*, or *orogenic movements*, and *epeirogenesis*, or *epeirogenic movements*, refer to concepts whose meaning is vague; these terms will be discussed later.

In view of these difficulties and the need for greater knowledge of the origin and mechanism of tectonic movements in order to classify them, the present author has meanwhile adopted the easiest solution of grouping tectonic movements according to the principal structures produced by them.

Three basic types of tectonic movement may be distinguished:

1. *Oscillatory movements*, manifested as upwarping and downwarping of the earth's crust and resulting in enormous structural uplifts and basins.

2. *Folding movements*, manifested as crumpling of the strata of the crust into folds.

3. *Rupturing movements*, manifested as fracturing of the crust.

This classification does not encompass cohesive deformations that do not result in folding: deformations of massive bodies, distortions, and internal plastic dislocations. These are tentatively regarded as special manifestations of folding.

The author realizes that the terms *folding* and *fracturing* movements are not quite exact: it would be more nearly correct to speak of *fold-forming* and *fracture-forming* movements. The less exact terms were chosen because brevity was preferred to scrupulous accuracy.

Magmatic phenomena are in a class by themselves. On the one hand, the formation of igneous bodies involves displacement of material that affects the structure of the crust. On the other hand, however, the specific nature of magmatic phenomena and the great role played in them by physicochemical processes make it necessary to distinguish them from tectonic movements proper, which are fundamentally mechanical. But whatever view is held of the relationship of magmatic phenomena to other geologic processes, they must under any circumstance be considered in connection with tectonic processes. This is due to the pro-

found and close interrelationships between the movements of the earth's crust and the movement of magma within it. Moreover, many features of tectonic activity and its basic principles cannot be understood unless magmatic processes are taken into consideration. This is particularly important in discussing the deep-seated causes of tectogenesis. Igneous activity is the only "peephole" through which one may glimpse the phenomena occurring at great depths below the earth's crust. For these reasons, the connection between igneous activity, tectonic movements, and the tectonic history of the earth's crust as a whole will be discussed specifically later. In the meantime the term *magmatic tectonic movements*, which was used earlier, will be discarded in favor of *igneous activity*, since it is more nearly correct to consider igneous activity as intermediate between tectonic processes proper, which have a purely mechanical character, and those processes, still completely unknown, which, occurring mainly below the crust (and to some extent within it), are essentially physicochemical and belong rather to the field of geochemistry than to geotectonics.

The three types of tectonic movement are briefly described below.

1. *Oscillatory movements* are vertical movements of the crust, including both uplift and subsidence: the crust is upwarped in some places and downwarped in others. The most important feature of these movements is the fact that in the same place uplift may be followed by subsidence, and vice versa. The direction of the movement is reversed; thus the movement is oscillatory. It should not, however, be imagined that after an uplift of a given part of the crust, this same part will subside in its original shape and area. In oscillatory movements the outlines, dimensions, and location of uplifted areas change continually, sometimes rapidly and sometimes slowly, and to a varying degree. Therefore successive uplifts and subsidences never repeat each other exactly, and the process as a whole cannot under any circumstances be regarded as oscillation of plates hinged at definite points. It is rather a continually changing wavelike process (Belousov, 1938a and 1948).

In current non-Soviet literature oscillatory movements are frequently called *epeirogenic*, and the movement as such is called *epeirogenesis*. These terms were introduced by Gilbert (1890) and until lately were universal. Among the more recent writers outside the U.S.S.R., Stille (1918) most clearly defined epeirogenic movements. The characteristic features of these movements, he believed, are their slowness and the fact that they do not alter the structure of the earth's crust. Neither idea is true: among definitely oscillatory movements there are rapid, short-lived uplifts causing gaps in the formation of sediments. It will also be seen that oscillatory movements causing uplifts and downwarps of the crust change its structure by creating large structural forms within it. The present author's definition of oscillatory movements seems to correspond more closely to actuality.

The word *epeiros* means "continent," and *epeirogenic* means literally "continent-making." But vertical movements of the crust are directed not only upward but downward as well; in the latter case they destroy continents and lead to the formation of seas in their stead. The term is thus definitely unsatisfactory.

Some writers have supposed that large-scale downward movements of the crust could be designated by the term *thalassogenic*, or "sea-forming" (*thalassa* means "sea" in ancient Greek). This suggestion is quite unacceptable because the upward and downward movement of the crust is a single process of oscillation, a single type of movement, which cannot be split into halves.

V. Ye. Khain (1939) proposed that two types of movement be distinguished: epeirogenic and oscillatory. By the first he meant the slow, protracted formation of large downwarps and uplifts of the crust—its wavelike movement—and by the second he meant rapid, short-lived pulsations (upward or downward), usually occurring with equal intensity over large areas and superimposed upon the movements of the first kind, complicating them. It will be seen later that this distinction corresponds to the observed facts.

2. *Folding movements* result in crumpling of strata into folds. Folding is a plastic deformation of the earth's crust.

Instead of folding movements, many writers speak of *orogenic movements* and *orogenesis*. The meaning of this term, however, is broad; it implies not only folding but also other dislocations such as faulting. Like "epeirogenesis," this term was introduced by Gilbert and achieved universal acceptance. It too must now be considered obsolete. According to Stille's definition (1918), *orogenesis* differs from *epeirogenesis* in that it proceeds rapidly by short-lived impulses and leads to changes in the structure of the crust. We have already seen that this is not the difference between the two types of movement. *Orogenesis* means literally "mountain building," and folding and faulting are often called *mountain-forming movements*. But it is known that mountains are not formed directly by these movements; they form as the result of vertical uplifts of the crust; that is, as the result of oscillatory, or, to use the old terminology, epeirogenic, movements. This leads to an absurdity: mountain-building movements do not form mountains, and epeirogenic movements do not create continents.

3. *Rupturing movements* produce cracks or fractures in the crust along which relative displacement of crustal blocks may or may not take place. In the latter case, the rocks are fractured but retain their original position. Tectonic movement in this case is at a minimum, consisting merely in widening of the fractures.

Rupturing movements are often assigned to one general orogenic type of movement, which is divided into two subtypes: *plicative* (folding) and *disjunctive* (faulting) movements. The term *orogenesis* has been

discussed. The term *disjunctive movement* can be translated as "rupture" and is thus identical with our term.

In speaking of large-scale displacement involving normal faults, some authors use the term *taphrogenesis* (*taphros* means "ditch"). This term does not embrace all rupturing movements as they are understood here.

All these types of tectonic movement develop together, in regular combinations and succession. The tectonic regime of a given area is never determined by a single type of crustal movement but is always manifested as a combination of various tectonic transformations. A geosyncline, for example, is typified by the combination of intensive movements of all types, whereas the general features of a platform are characterized by the low intensity of these various movements.

To interpret correctly the observed combinations of tectonic movements determining the complex tectonic character of a given region, one must first of all study the nature, mechanism, and principles of the development of each type of tectonic movement separately. In the following sections we shall first consider, in succession, the problems pertaining to oscillatory, folding, and rupturing movements. Thereafter we shall discuss the relationship between tectonic movements and igneous activity, and, finally, we shall turn to the different combinations of types of tectonic movement that may be observed and the general principles of the tectonic process as a whole.

This survey may begin with oscillatory movements as the basic form of geotectonic phenomena.

It is almost impossible to observe tectonic movements directly. The only exceptions are present-day manifestations of oscillatory movements of the crust and earthquakes reflecting movements along tectonic faults. In the vast majority of cases one may observe only the results of tectonic movements as indicated by the various occurrences of rocks, from which one must reconstruct the history of past movements.

Present-day oscillatory movements are discussed in the next chapter. In the following chapters movements that have occurred in past geologic epochs are discussed and much attention is given to the methods of analyzing structures in terms of the mechanism and history of their formation.

PART FOUR

**OSCILLATORY TECTONIC
MOVEMENT**

CHAPTER 11

PRESENT-DAY AND RECENT OSCILLATORY MOVEMENT

Present-day crustal movements are those which are taking place at the present time, before our eyes, or have occurred in historic times. Recent movements are those which have occurred during the Quaternary period.

MANIFESTATIONS OF PRESENT-DAY OSCILLATORY MOVEMENT

Even in ancient times it was observed that seashores do not remain in one place but change position: either the land increases its area at the expense of the sea, or the sea invades the land. These changes are slow, but noticeable over periods of a few hundred years. Many observations were accumulated in later times that made it certain that seashores do not, indeed, remain immobile (Mushketov, 1935; Nikolaev, 1949).

In 1620, a harbor, Torne, was built on the shore of the Gulf of Bothnia. Less than a hundred years later it became useless because of the shallowing of the sea. In 1724, when this region was visited by Celsius, many old landings were far from shore, and where large ships could sail before, even small boats could make their way only with difficulty. Inhabitants of the seashore villages of northern Scandinavia, especially fishermen who know the depth of water near the shore, have, over the centuries, marked slow but noticeable shallowing of the sea. Mariners cannot use old maps of the Åland Islands because the depth of the sea has decreased everywhere in this vicinity during the past few decades.

At first the retreat of the sea from Scandinavian shores was explained by a decrease in the total amount of water in the world's oceans. But in that case shallowing of the sea should occur everywhere, since all the seas are connected. Meanwhile, during the same few decades, the opposite process, rising of the sea level and submergence of the land, was observed elsewhere. Thus in Ravenna, Italy, near the mouth of the

Po on the Adriatic shore, there is an old roadway that is now below sea level. During historic times a strip of shore with ancient Roman buildings near Naples in southern Italy has subsided, and the buildings are now submerged. The shores of Belgium, and especially of Holland, are subsiding, and the inhabitants must continually protect their lands by building dikes.

The simultaneous recession of the sea in one place and advance in another clearly indicate that the basic cause of this phenomenon is the movement of the lithosphere: in Scandinavia the earth's crust is rising; in Italy, Belgium, and Holland it is subsiding. A general fluctuation of the sea level is not impossible, however, and the observed changes in shore lines may be due to the combined effect of rising or lowering of the sea level and of local movements of the lithosphere. World-wide fluctuations of sea level, manifested in a simultaneous and equal rise or fall of all waters, are called *eustatic fluctuations*. It is difficult to evaluate the relative importance of each of these factors.

These fluctuations may have different causes. One of them may be an actual change of the amount of water in the ocean. It is believed, for instance, that in times of extensive glaciation the amount of water in the ocean was diminished, since a considerable mass of water was locked on land in the form of ice. As the ice sheets melt, the amount of water in the oceans would increase. Calculations show that the melting of Quaternary ice sheets must have caused the sea level to rise by 60 to 100 m. On the other hand, similar fluctuations may result from changes in the volume of the ocean basins caused either by filling of the basins with sediment or by tectonic movements changing the crust. It was believed by Suess that changes in shore lines were caused by changes in the volume of the ocean basins. In his view, as the ocean basin becomes filled with sediments, the level of the ocean rises, the oceans become overfull, and transgressions result. Tectonic subsidence of the ocean bottom, on the other hand, increases the volume of the basin, the ocean level is lowered, and a regression occurs (Suess, 1883-1909).

It may be stated, on the basis of modern data, that the movement of shore lines is caused primarily by relative vertical movements of individual parts of the earth's crust, its rise and subsidence occurring simultaneously in different places but in different directions and at different rates. Nevertheless, the significance of the fluctuations of the ocean itself cannot be completely denied. The filling of the basins with sediments must be considered a factor of very little importance, since the accumulation of sediments occurs simultaneously with subsidence of the crust, so that there is little change in the depth of the sea. The changes in the amount of water in the ocean must have been relatively small in recent geologic periods. There remains the effect of tectonic activity:

the general change in the volume of the ocean basins as a result of vertical crustal movements. The same applies to shore-line fluctuations. For the sake of simplicity, it may be considered that the movements of shore lines observed in modern times are entirely the effect of rising and subsidence of the lithosphere, since the differences in direction of movement of the shore lines in different places indicate that local uplifts and subsidences of the land play the principal role in all these phenomena.

In a number of cases, observations of shore-line movements occurring in historic times have shown that they are oscillatory—that the direction of movement in a given place changes with time. On the shore of the Bay of Naples (Lyell, 1866), a little to the north of the small town of Pozzuoli, there are ruins of a Roman temple. Among the ruins stand three marble columns, which were unknown until the middle of the eighteenth century. In 1749 their tops were discovered among the bushes on the seashore, rising only a few feet above the ground. The greater part of the columns was buried in soil. They were excavated and found to be about 12 m high. Beneath them was a marble floor. To the height of 3.6 m above the pedestals the columns were perfectly smooth, but above this there was a 2.7-m band showing borings; above this height they are again smooth. Clearly there was a time when the columns were submerged under the sea to a depth of 6.3 m. Their lower parts were protected by rubbish, sediment, volcanic ash, and deposits of a nearby hot spring (all these materials were found during the excavation), and were protected against the rock-boring mollusks. The upper parts rose above the sea, and only the 2.7-m band was in the water. Since the temple had been built on land, it is obvious that in later times this part of the shore was submerged to depths of more than 6 m. But in the eighteenth century the columns were once more above sea level, so an uplift must clearly have followed the subsidence. Historical records make it possible to refine this chronology. The temple was built about 200 B.C. The period of maximum submergence antedates the end of the fifteenth century. In the middle of the sixteenth century the region began to be uplifted and remained at high level until the end of the eighteenth century. In the beginning of the nineteenth century subsidence began again. In 1828, according to Lyell, the pedestals of the columns were 1 ft below sea level. According to another observer, the subsidence between 1822 and 1838 was at the rate of 7 mm per year. It continued during the entire nineteenth century and is still in progress: in 1878 the base of the columns was 65 cm below sea level; in 1911, 188 cm; and now it is more than 2 m below sea level.

A similar case is known in Sweden (Lyell, 1866). There, during the digging of a canal in 1819 approximately 25 km south of Stockholm,

marine sediments were discovered containing shells of a modern Baltic species. At the depth of 18 m a fisherman's hut was found buried in the sediments. Evidently, after the building of the hut, the region subsided and was covered by the sea. The hut was buried in the sediments. Later an uplift began and is still continuing. Such cases justify the term *oscillatory movements* as applied to modern uplifts and subsidences of the crust.

We have discussed only those present-day movements that are manifested in fluctuations of shore lines. Sometimes changes in the elevation of points distant from the sea are observed. Usually this is noticed because of changes in visibility: for example, the top of a tower, belfry, etc., which was visible from a certain point, becomes invisible, or some object formerly invisible rises above the horizon. Such observations, however, are rare, and it is difficult to ascertain the existence of slow crustal movements on land where there is no datum from which measurements could be made. This can be done only by precise methods, which will be described below.

METHODS OF STUDYING CURRENT OSCILLATORY MOVEMENTS AND SOME RESULTS OF THEIR APPLICATION

Many different methods are used in investigating present-day and recent crustal movements. Some of these produce rough qualitative results; others reveal the history of the movements with great precision. These methods can be classified as historical, geologic, geomorphologic, and geodetic (Nikolaev, 1949).

The *historical method* has already been illustrated. It includes various ways of establishing oscillatory movements by the change in the position of buildings relative to the seashore, historical records, archeological data, and so forth. This method is useful in revealing crustal movements that have taken place during the period of human culture and is restricted essentially to seashore regions; its exactness varies greatly in different cases.

Geodetic methods provide the most precise means of studying modern oscillations of the earth's crust, and they have been used widely in recent years. These consist essentially of repeated precise leveling along definite traverses. Comparison of leveling done decades apart permits precise reconstruction of changes in the relative and absolute elevations of individual points.

To determine sea level fluctuations, special gages are installed along seashores. Several such gages were set up in Scandinavia many years ago. The records of two of them, installed on the shore of the Gulf of Bothnia at Pitea and Ratan, are shown in Fig. 77. At Pitea recordings began in 1649. The curve shows that from then until 1908 there was a

uniform lowering of the sea level relative to the land totaling 265.4 cm; that is, the land rose at the average rate of 1.02 cm per year. A similar record is seen at Ratan.

Among the projects of precise leveling on land, one of the most interesting is the recent work of Yu. A. Meshcheryakov (1958) done within the Russian platform. Comparison of the results of leveling done in 1913 to 1932 and 1945 to 1950 has shown that the average rates of uplift and subsidence on the Russian platform are 2 to 4 mm per year. This analysis revealed uplift in the vicinities of Tallin (+2.3 mm per year), Vilnius (+3.8 mm per year), Kursk (+3.6 mm per year), Khar'kov (+3.9 mm per year), and the area of the Donbas (+3.7 mm per

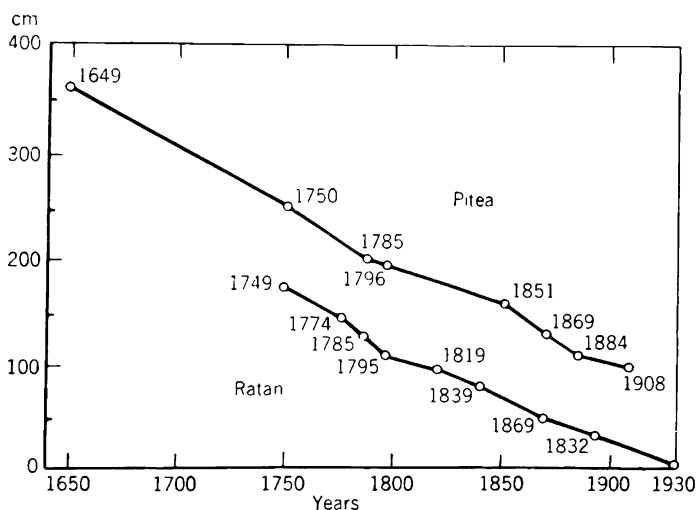


FIG. 77. Sea level changes at the towns of Pitea and Ratan (Sweden). The years of the observations are marked along the curves (after C. Tsuboi).

year) and subsidence near Vitebsk (−1.4 mm per year), Moscow (−3.7 mm per year), Leningrad (−3.8 mm per year), and Odessa (−5.1 mm per year).

Figure 78 shows the vertical displacement of points on a cross section of the main island of Japan during the period from 1896 to 1928. All points of the first leveling are on the zero abscissa. The diagram shows the relative vertical displacements of individual points that have occurred between the first and second leveling. It will be seen that almost all the points have been lowered, but less so in the middle of the island than on its margins. The amount of lowering varies from a few millimeters to 9 cm. The surface of the island has thus been upwarped like a dome.

The results of four repeated levelings made in Japan along the line

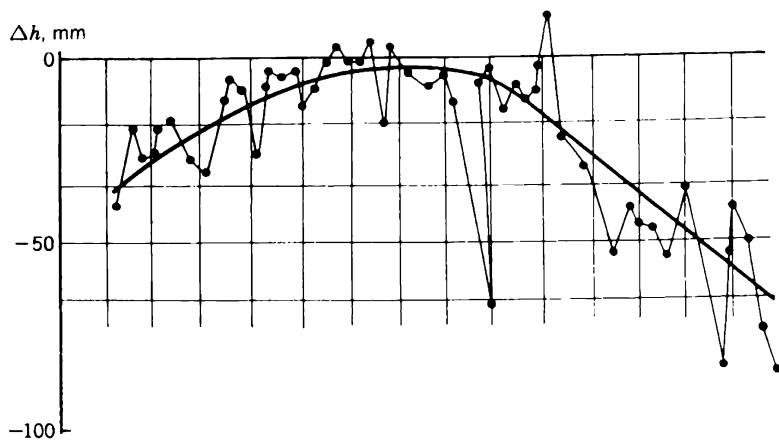


FIG. 78. Change in elevation along the line of precise leveling from the Inland Sea to the Sea of Japan (Japan) in the period from 1896 to 1928 (after C. Tsuboi).

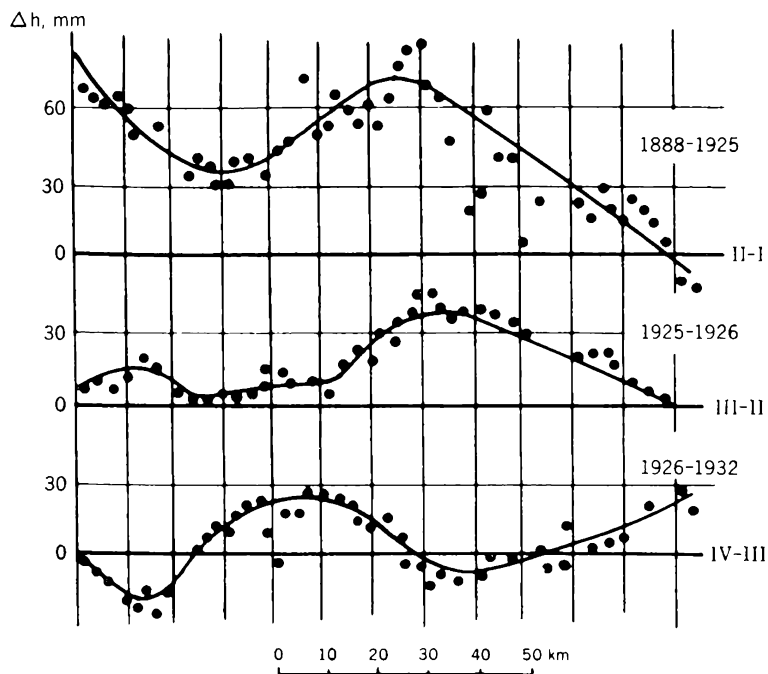


FIG. 79. Changes in elevation along the line from Takasaki to Shimosuwa (Japan) from 1888 to 1932 (after C. Tsuboi).

from Takasaki to Shimosuwa during the period from 1888 to 1932 are shown in Fig. 79. Each successive curve indicates the relative vertical movement of points during the time elapsed since the preceding leveling; that is, the line of elevations obtained during the preceding leveling is regarded as the zero line in constructing the next succeeding diagram. The curves show that between first and third levelings (1888 to 1926) the crust moved in the same direction at each of the points. Between the third and fourth levelings (between 1926 and 1932), the direction of movement changed, and the majority of points previously rising subsided slightly, and vice versa. The rate of movement is in millimeters per year (Tsuboi and Chuji, 1939).

It should be noted that there is no absolute certainty that all the figures cited above are correct, since in many cases the precision of measurement is less than the recorded changes. Again, all leveling is referred to sea level as the datum, with the assumption that it remains constant. But if this condition is not fulfilled, and the sea level fluctuates, the changes cannot be detected and the eustatic change in the sea level enters into the results just as it did into the results of observations made along the seashore.

A method of studying present-day crustal movements based on observations of changes in the level of lakes was used by G. Yu. Vereshchagin (1926, 1931). If the earth's crust is tilted in any direction as the result of nonuniform uplift or subsidence, then the water in the lake must flow toward one shore and recede from the other. Vereshchagin made observations on Lakes Segozero (Karelia) and Onega; he found that the southeastern shores of both lakes are subsiding while the northwestern shores are rising. Using the same method, Freeman, in America, was able to show by his observations of changes in the water level of the Great Lakes that the region is being tilted about a horizontal axis running $N72^{\circ}W$, with its northern end rising and its southern end subsiding. The rate of tilting is such that in 100 years an additional tilt of 10 cm/100 km is produced.

The rate of tilting of the land can be determined directly by special instruments called *tiltmeters*, constructed on the principle of sensitive seismographs. Using this method, M. Schmidt discovered that the surface of France during the period from 1857 to 1884 was tilted toward the southwest at the rate of 40 cm/100 km/100 yr.

The results of geodetic measurements covering a certain area may be shown graphically by lines of equal uplift (isoanabases) or of equal subsidence (isokatabases). These lines connect points rising or subsiding at the same rate or points of equal displacement over a given period of time. As an example, an isokatabase map of France is reproduced in Fig. 80, which shows the amount of subsidence in centimeters during a twenty-seven-year period (1857 to 1884).

Geomorphologic and geologic methods must always be used together. The methods already described reveal oscillatory movements that are going on now or have occurred during the past few centuries, but geomorphologic and geologic methods make it possible to judge of these movements in the perspective of much longer periods of time, embracing all the Quaternary period or a considerable part of it. These methods are used to study not current movements (as we have defined them) but recent ones. They are based on modern geomorphology, which by the analysis of relief forms establishes the direction of present or recent vertical movement of a given region, and also on geologic data, which

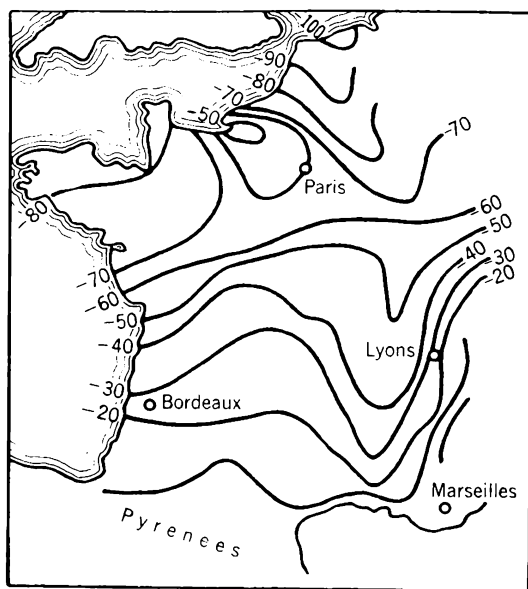


FIG. 80. Isokatabase map of France, 1857 to 1884. Numbers indicate subsidence in centimeters (after M. Schmidt).

reveal the evidence of such movements by the attitude and composition of rocks.

The traces of old shore lines are used along the seashore. They may be terraces uplifted above the present sea level, wave-cut cliffs and notches, bars, or sediments containing modern marine shells. These phenomena indicate the uplift of the land. Often these features of ancient shore lines are found at several levels, indicating several stages of uplift. The latter, in this case, must have occurred discontinuously, now accelerating, now stopping, so that there was time for wave action to create, at a given level, either a wave-cut platform or a bar.

In Scandinavia ancient shore lines are found at elevations of 5 to 200 m above the present sea level. In tracing individual terraces or

wave-cut cliffs it has been observed that a given terrace has different elevations at different places, thus indicating a differential uplift of the earth's crust. By recording the number of terraces and other features of ancient shore lines and by studying their elevations in different places, it is possible to determine how the uplift was distributed in a given territory and how this distribution changed with time.

In this way an isoanabase map of Scandinavia was drawn from the elevations of outcrops of the Yoldia Sea deposits (Fig. 81). These occur at elevations of 0 to 200 m or more above the present sea level. Originally they were, of course, at the same level, and their present distribution indicates nonuniform rising of the land. The shape of the

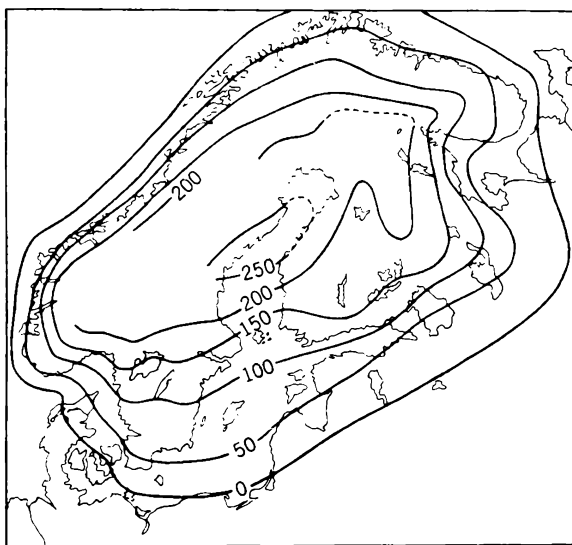


FIG. 81. Isoanabase map of Scandinavia: lines of equal elevation of the bottom of the Yoldia Sea are marked in meters (*after Høgbom*).

isoanabases shows that the maximum uplift occurred in the northern part of the Gulf of Bothnia, decreasing from there in all directions. Since Yoldia Sea time Scandinavia has been elevated as a great dome.

These methods are more successful in establishing uplifts than they are in establishing subsidences, for in the latter case the old strand lines with their complex of relief forms are covered by the sea. For example, it can be observed at the mouth of the northern Dvina River that a river terrace covered with peat bogs is being gradually lowered below sea level. Sometimes submerged terraces can be discovered by measuring the depth of the sea; such measurements may also disclose other features of subaerial relief on the bottom of the sea, such as the continuation of stream valleys (Klenova, 1948). A number of rivers falling into the sea continue on its bottom as winding channels. This

is true of the rivers on the French Atlantic seaboard (the Seine, Adour, Chatolin, etc.), and of some American rivers (the Hudson). The rivers Elbe, Weser, Rhine, and Maas have very long submarine extensions. The channels of the first three can be traced all the way across the North Sea. At one time these rivers and the stream that flowed on the present site of the Skagerrak joined together and emptied into the ocean

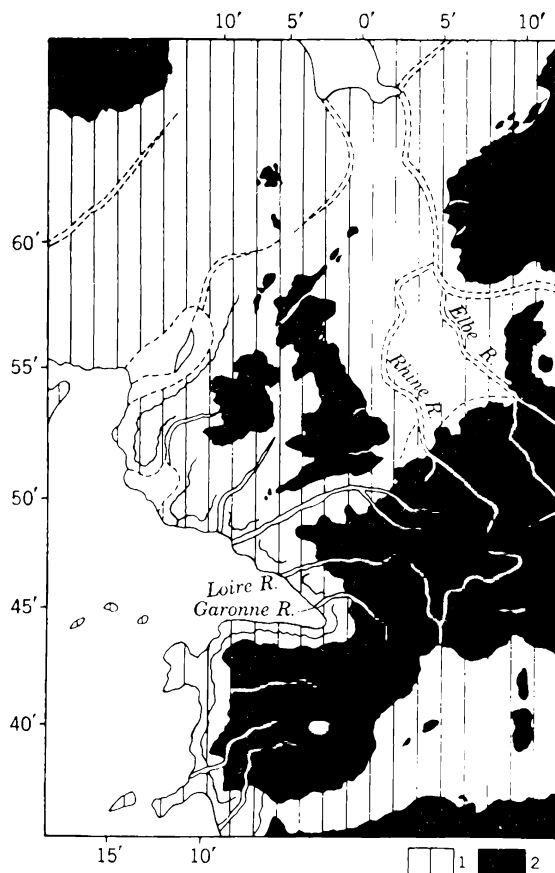


FIG. 82. The old river systems of northwestern Europe now submerged under the sea: (1) ancient land, now inundated; (2) modern land.

north of Scotland. At that time England was part of the continent, and the Seine flowed into the Atlantic where the English Channel is now. The rivers of the Atlantic seaboard of England can also be traced for great distances on the bottom of the sea to the very edge of the continental shelf (Fig. 82). The large estuaries at the river mouths (for example, the Gulf of the Ob') also were formed by submergence of the land and flooding of the valleys.

The problem of submarine canyons has recently received much attention. These are deep narrow canyons discovered in the past few years during investigations of the ocean bottom, on almost all the continental slopes (Klenova, 1948; Lindberg, 1947; Panov, 1948; Shepard, 1948; and others). Beginning at the junction between the shelf and the continental slope, they continue down the slope to the depth of three or more kilometers. Some writers believe that they were formed by sub-aerial erosion and conclude that in very recent times (during the last Ice Age) the level of water in the sea was several kilometers below the present level, so that the continental slopes were exposed.

This interpretation of the origin of submarine canyons is very improbable. If it were correct, half the water in the ocean would have had to disappear somewhere for a short time and then return. It is more probable that submarine canyons are the result of subaqueous erosion by density currents carrying suspended particles of silt and flowing down the continental slopes.

The study of marine sediments, not only those elevated above the present sea level but also those on the sea bottom, is a geologic method of investigation of the movements of the earth's crust that yields very important and interesting results. The Quaternary history of the Black Sea was deciphered with this method by A. D. Arkhangel'skiy and N. M. Strakhov (1938), who have distinguished four stages in the most recent history of the Black Sea. Marine sediments of the first stage (Old Euxenian) are found along the shore at elevations of 18 to 50 m above the present sea level. The second stage (Tyrrhenian or Karangatian) is recorded in marine sediments locally preserved at elevations of 1.3 to 25 m above sea level. Deposits of the third stage (New Euxenian) do not crop out on the present shore line and have been found only by oceanographic sounding of the sea bottom. The fourth stage is the contemporary Black Sea. Correlation of these data shows that, beginning with the Old Euxenian and up to the New Euxenian stage, the land rose relative to the sea but that later a certain amount of subsidence occurred and that the present sea occupies a greater area than that of the New Euxenian basin.

On the continents, recent oscillatory movements of the crust are established on the basis of the study of terraces, ancient erosional surfaces, transverse and longitudinal profiles of valleys, Quaternary sediments, etc. Since the methods of interpreting this evidence are discussed in works on geomorphology, this discussion will be limited to a few brief remarks.

The tiers of terraces found in river valleys, especially in mountainous regions, testify to lowering of the base level or to relative elevation of the land. The uplifts occur nonuniformly in time, sometimes at a rapid rate, and a channel is incised, sometimes slowly; then lateral erosion

begins, and a broad valley bottom is developed on the consecutive erosional surface. If the terrace accumulates sediment, it must be concluded that the uplift of the region has been interrupted by the opposite movement—subsidence—since erosion tends to produce a profile of equilibrium and deposition is possible only when the channel is lowered below this profile. Thus the entire process of formation of terrace steps in valleys proceeds under conditions of oscillatory movement, alternating uplifts and subsidences, with a general predominance of the former. The phenomenon of *terrace intersection*, however, indicates a temporary or local predominance of subsidence.

Tracing of terraces along a valley shows that uplifts are not uniform in area. The terraces in mountain ranges have a greater slope away from the axis of the range toward the foothills than the slope of the normal profile. Moreover, toward the heads of valleys the terraces, in cross section, spread apart as a rule, and the distances between them increase. The number of terraces also increases in that direction, and they appear to branch out. This phenomenon, illustrated by actual examples in texts on geomorphology, indicates that mountain ranges are formed as the result of arching of the earth's crust. The movement is more intensive in the axial zone of the range and dies out toward its margins, where subsidence often replaces uplift and the terraces are lowered beneath the present level of the land and covered with sediments.

Old erosional levels or surfaces and ancient uplifted peneplains found in mountainous regions at different elevations have the same significance as the terraces. These surfaces of ancient planation, dissected by valleys and preserved as flat remnants on the summits and slopes of mountains, are parts of a domelike surface produced by up-arching of the range. The highest of all remnants of ancient planation are at the level of the summit. The surface that connects all the peaks of a given mountain system is usually a more or less regular dome, high in the middle and sloping gradually toward the margins.

All these phenomena lead to the conclusion that mountain ranges are the result of prolonged archlike uplift divided into epochs of relatively rapid movement and epochs of standstill or even slight subsidence. The uplift is accompanied by dissection of the region into elevated inter-stream divides and deep valleys forming a complex incised mountain relief.

DISTRIBUTION OF PRESENT-DAY AND RECENT OSCILLATORY MOVEMENTS

By using all the methods enumerated above, it has been possible to construct maps of the distribution of present-day and recent oscillatory

movements of many regions of the earth. One of the most complete maps of this type was drawn for the territory of the U.S.S.R. by N. I. Nikolayev (1949); it is shown in simplified form in Fig. 83. The following regions have been outlined on the map (in order of the numbers on the legend):

1. Regions of intensive Quaternary and present-day uparching (Fennoscandia, Taymyr, and the Novosibirsk Islands).

2. Regions of intensive Quaternary uplifts (elongated along the strike) with local linear subsidences characterized by high gradients, numerous fractures, and renewed movements on ancient faults (Caucasus, Tien Shan, eastern Sayan, and Stanovoy Mountains).

3. Regions of weaker Quaternary and present-day linear uplifts and subsidences with considerable gradients, fracturing, and renewed movements on ancient faults (Carpathians, western Sayan, Kopet Dag, Far East, and northeastern U.S.S.R.).

4. Regions of weak present-day and Quaternary uplifts and subsidences, with weak renewed movements on ancient faults (Urals, Mugodzhaz Mountains, Ulutau Mountains, and the Chingiz Range).

5. Regions of poorly differentiated Quaternary movements, predominantly positive (central Siberian plateau and the Donbas).

6. Regions of poorly differentiated linear movements, with weak positive movements predominating (northeastern Kazakhstan, eastern slope of the southern Urals, Polar Urals, and the Turanian depression in the Amu-Daria region).

7. Regions of weak Quaternary and present-day positive movements (Turgai upland, southern part of the Russian platform, and Ust'urt).

8. Regions of the platform marked by uplifts increasing in intensity in the second half of the Quaternary period (Polish basin and Moscow basin).

9. Regions of compensated uplift and subsidence in Neogene and Quaternary times. Weak uplifts occurred during the latter (west Siberian plain and individual areas in the south of the European part of the U.S.S.R.).

10. Regions of noticeable subsidence and differential movements (northern part of the west Siberian plain and the lower waters of the Vilyuy, Amur, and Lena Rivers).

11. Regions of dominant Quaternary subsidence (Kolyma-Yana plain, Khatanga depression, and Kyzyl-Kum).

12. Regions of intensive Quaternary downwarping with large gradients (Black Sea and Caspian depressions; Kara-Kum, Balkhash, and Alakul depressions).

13. Regions of intermontane basins with clearly defined Neogene, Quaternary, and present-day downwarping (Rion, Kura, Ferghana, Issyk-Kul', Baykal, and Muysk depressions).

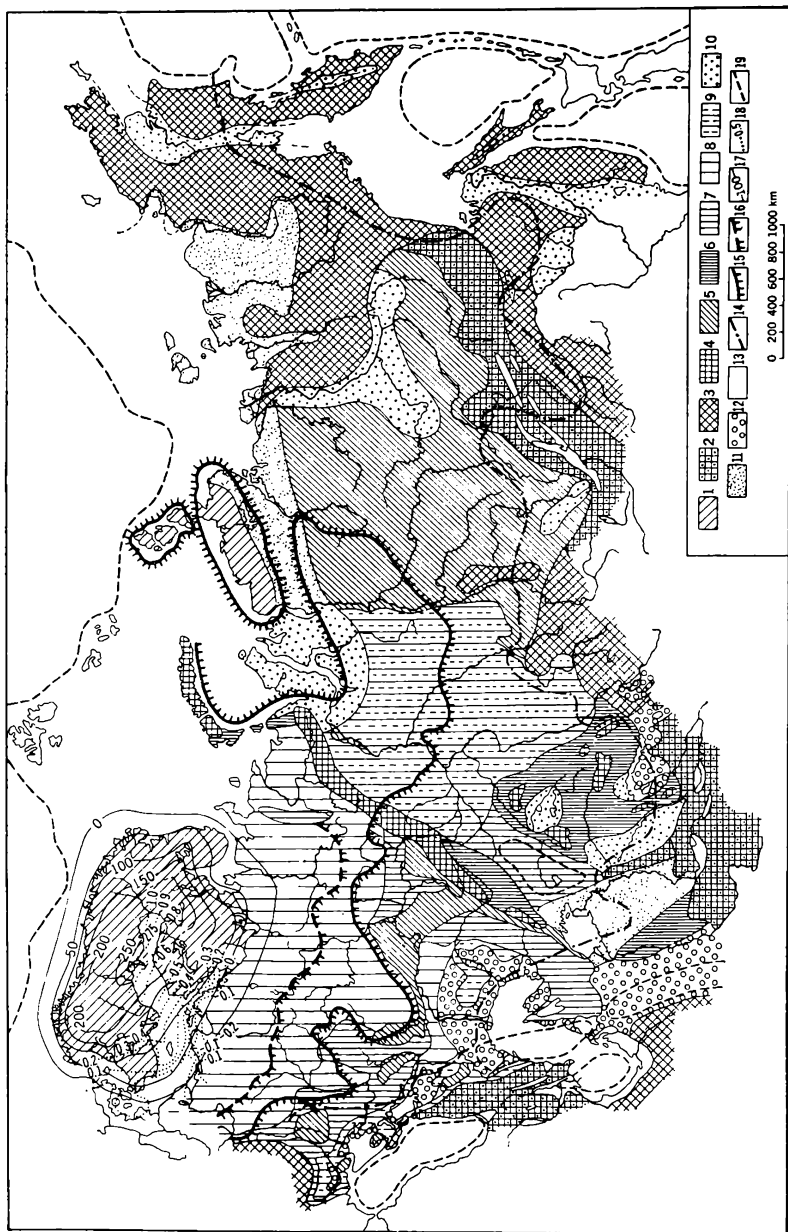


FIG. 83. Recent oscillatory movements in the territory of the U.S.S.R. (explanation in the text) (after N. I. Niko-
layev).

In addition, the map shows the following:

14. The northern limit of earthquakes of V degree of intensity.

15-16. Approximate limits of Pleistocene glaciation in the western U.S.S.R. (15. Riss, 16. Würm).

17. Contours showing late Pleistocene and post-Pleistocene uplifts in Scandinavia (in meters).

18. Contours of present-day movements in Scandinavia constructed on the basis of instrumental data (in centimeters).

19. Contours showing the observed relative uplift and subsidence on the continent and in the sea.

It can be seen from Fig. 83 that the U.S.S.R. is divided into many regions, undergoing crustal movements in different directions and at different rates. The mountainous areas are everywhere regions of uplift, whereas plains and depressions are more often regions of subsidence. This reflects the intimate relationship between the relief and recent oscillatory movements: mountains are formed by crustal uplift with simultaneous dissection by stream valleys. The necessary condition for peneplanation is cessation of uplift, so that the erosion has time to carve out wide valleys and smooth out the relief. For a perfect plain to come into being, subsidence is necessary, because then streams are slowed down and great masses of sediment are deposited, filling depressions and masking relief.

The zone of intensive uplift, corresponding to the highest young mountains, encloses the Soviet Union in a wide arc extending from the south and east. Within this arc lie regions of lesser uplift or subsidence. Thus our country has the form of a vast amphitheatre sloping to the north.

In the U.S.S.R., present-day oscillatory movements are most clearly manifested in the Tien Shan Mountains. Here many late Tertiary and early Quaternary erosional surfaces are found that have been strongly deformed by the still active upwarping of the mountain ranges and downwarping of the intermontane basins. Locally, these erosional surfaces have acquired dips of 5 to 20°, forming domes within the ranges and troughs in the basins. The amplitude of uplift and subsidence in very late geologic time is up to several kilometers (Shul'ts, 1948; Krestnikov, 1954). To a greater or less extent, similar phenomena are observed in all mountainous countries: elevated and warped relatively young erosional surfaces and river terraces testify to the considerable amplitude of Quaternary oscillatory movements.

Evidence of a recent uplift is very clear on the Murmansk coast, where at an elevation of 90 m above the present sea level one can see an ancient wave-cut bench. The region of the lower reaches of the northern Dvina and Cheshskaya Bay, on the other hand, is subsiding. On the Apsheron peninsula, in the vicinity of Baku, historical data in-

dicating one complete oscillation (uplift and subsidence) with an amplitude of 16 mm in the past 800 years. An inn built in the twelfth century on the shore of the Bay of Baku was completely submerged by the sea and later partially raised above it. Subsidence is taking place at the mouth of the Volga and in the entire Caspian depression.

In Europe, present-day and recent oscillatory movements have been especially well studied in Scandinavia. Tidal gage records have made it possible to determine the absolute rate of current uplift at many points. Uplift is greatest in the north of the Gulf of Bothnia, decreases gradually southward, and is finally replaced by subsidence. A few figures characterizing the rate of present-day oscillatory movements in Scandinavia and adjacent regions are given in Table 11.

*Table 11. Rates of Present-day Crustal Movements in Scandinavia
(After Tsuboi)*

<i>Locality</i>	<i>Uplift (+) or subsidence (-), mm/yr</i>
Ratan.....	+10
Vastervik.....	+ 4
Kronstadt.....	+ 3
Klaypeda.....	+ 1
Copenhagen.....	- 0.65

Holland and Belgium are subsiding predominantly along the sea-shore. The subsidence of the shores of Holland during a century (1800 to 1900) amounts to 30 cm. Southern England is also subsiding, whereas Scotland is rising. The Alps are also rising. Thus there is a zone along the Baltic and the North Seas that is undergoing downwarping and is bounded on the north and south by regions of uplift.

The crustal movements in the Mediterranean region have already been mentioned. It may be added that the mouth of the Tagus is subsiding, whereas Gibraltar, the Balearic Islands, and Malta are rising at the rate of 1 cm per year. Crete is slowly being tilted: its western end is rising and its eastern end subsiding.

No less varied are the present-day crustal movements of North America. Along the Atlantic coast, Labrador is rising, together with large adjacent areas of Canada; the uplift diminishes toward the south and changes to subsidence. Uplift predominates on the Pacific coast of North America. The gradual rising of the Pacific shores of South America was noted by Charles Darwin.

The ocean floors, like the continents, undergo vertical oscillations. There is no doubt about the littoral zone, which obviously moves with the adjacent land areas. But even on islands far away from the mainland there are terraces indicating movement of land with respect to the sea. In the middle of the last century Darwin advanced his hypothesis of

the origin of coral reefs, postulating that the reefs grow upward, reaching for the surface of the water as the earth's crust is downwarped. This theory has been confirmed by drilling on Funafuti and the Bermuda Islands, where the boreholes at the depth of approximately 300 m below sea level were still in the coral structure. Since corals live at

Table 12. Rate of Present-day Crustal Movement at Certain Points

<i>Locality</i>	<i>Rate of uplift (+) or subsidence (-), mm/yr</i>
Canada:	
Quebec...	+0.9
Halifax.....	-4.3
U.S.A.:	
Boston.....	-4.5
New York.....	-4.2
Charleston..	-5.7
Miami.....	-4.8
Seattle.....	-1.3
San Francisco.....	-1.8
La Jolla.....	-1.4
Honolulu.....	-2.1
Argentine:	
Buenos Aires.....	-1.1
La Plata.....	-10.0
Philippine Islands:	
Manila.....	-1.4
Japan:	
Isyoro.....	-1.0
Abaratubo.....	-1.4
United Kingdom:	
Aberdeen.....	+0.5
Liverpool.....	-0.7
Egypt:	
Port Said.....	-4.8
Australia:	
Sydney.....	+0.4
Burma:	
Rangoon....	-0.1
India:	
Bombay.....	-0.2
Madras...	-1.2

depths not exceeding 40 m, the growth of an atoll can occur only by simultaneous gradual lowering of the ocean bottom. But since at present the surface of the coral reefs is above sea level and corals cannot live above water, it is evident that downwarping must have been succeeded by uplift. Once again the oscillatory character of crustal movements is in evidence. Even more interesting data on crustal movements in the oceans have recently been obtained by drilling at Eniwetok Atoll.

Thus there is no reason to suppose that contemporary oscillatory movements on the ocean bottom differ from those on land. As on the continents, these movements create relief on the ocean floor.

THE IMPORTANCE OF STUDYING CURRENT OSCILLATORY MOVEMENTS OF THE CRUST

The study of present-day crustal movements is of great *practical importance*. The rising and lowering of the earth's surface cause changes in drainage, affect the intensity of stream flow, and in some cases may change the direction of flow and lead to the reordering of an entire drainage system. Such changes are slow and do not produce significant results within a single human lifetime. But man builds large structures designed to last for longer periods of time, such as dams, hydroelectric stations, and irrigation systems. Oscillatory movements cause the shore lines to shift and must therefore be considered in the building of harbor structures. Present-day crustal movements must obviously be kept in mind if such structures are to endure.

These movements also cause relative vertical displacements of triangulation points, change the geodetic net, and must be reckoned with in geodetic work.

Investigation of present-day movements of the crust is also of *theoretical importance*, since these movements represent almost the only tectonic process that can be observed and measured directly. With the aid of these observations, one can better understand the crustal movements of the past geologic periods.

Because of the great practical and theoretical significance of investigations of contemporary oscillatory movements, they are now intensively pursued in the Soviet Union. They have acquired a special importance in connection with the colossal hydro constructions that are being built there.

CHAPTER 12

OSCILLATORY MOVEMENTS IN PAST GEOLOGIC PERIODS

EVIDENCE OF OSCILLATORY MOVEMENTS IN A GEOLOGIC SECTION

Studies of geologic sections indicate that there were oscillatory movements of the earth's crust in past geologic epochs, as well as in the present.

As an example, we may take the section through the Middle and Upper Carboniferous coal measures of the Donbas. It is well known that this series is composed of closely interlayered, comparatively thin beds of sandstone, shale, limestone, and coal. Paralic deposits of coal are formed at sea level, in swampy lagoons. The rocks surrounding the coals contain shallow-water marine fauna; they were deposited at depths of no more than a few score meters. Thus the changes in the rocks composing the section testify that during the Middle or Upper Carboniferous in this region there were oscillatory movements of the earth's crust that elevated it to the level of water or caused it to sink to a not very great depth (Fig. 84).

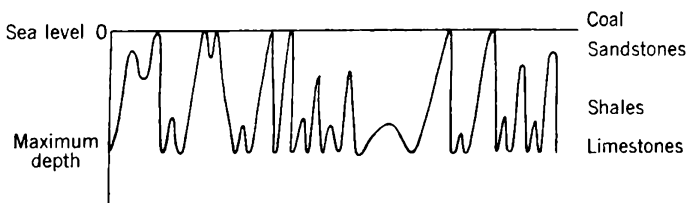


FIG. 84. Oscillatory movements of the earth's surface in the Donbas during the C₂ epoch (after P. I. Stepanov).

An examination of this section will also lead to another conclusion. The Middle and Upper Carboniferous of the Donbas reaches a great thickness. According to some data, its thickness is as much as 10 km; in any case, it is not less than 6.5 km. Throughout all this thickness, the section is generally composed of a similar complex of shallow-water sediments. In this respect there are no important differences between

the lower and upper parts of the section. It is quite clear that such a great thickness of uniformly interbedded shallow-water deposits could have been formed only under conditions of gradual subsidence of the earth's crust simultaneous with the accumulation of sediments. There was no earlier depression that was later filled with sediments.

The entire process took place in such a manner that the earth's crust subsided and the depression thus formed was simultaneously and continuously filled with sediments. As a result, the surface of the sediments always remained near the water level, and notwithstanding the subsidence of the crust, the basin preserved its shallow-water character.

Thus, in the first place, the change in the facies of the rocks from bed to bed has made it possible to establish the presence of oscillatory movements of the earth's crust, or (to be more exact) of the surface of sedimentation; in the second place, from the great thickness of the sediments one may deduce the considerable subsidence of this surface.

GEOMORPHOLOGIC REFLECTION OF OSCILLATORY MOVEMENTS

Let us imagine a part of the earth's crust in which there is subsidence relative to the adjoining parts and accumulation of the sediments proceeding simultaneously with the subsidence. There are evidently three possibilities:

1. The rate of accumulation of sediments is lower than the rate of subsidence of the crust.
2. The rate of sedimentation is equal to the rate of subsidence.
3. The rate of sedimentation is greater than the rate of subsidence.

In the first case, the surface of the lithosphere, or the surface of the continuously forming deposits, will be sinking and the depth of the basin will increase, but more slowly than the subsidence of the earth's crust. In the second case, the surface of the lithosphere will not change its hypsometric position, the depth of the basin will remain unchanged, and no movement of the earth's surface will be observed, since the sinking of the crust will be entirely compensated by the accumulation of sediments. Finally, in the last case, inasmuch as the accumulation of sediments is more rapid than the subsidence, the surface of the lithosphere will rise and the basin will become shallower in spite of the subsidence of the earth's crust.

Let us now consider the regions in which the earth's crust is rising. As soon as the surface of the earth is elevated above sea level, the forces of denudation, which tend to lower the region, become active. Therefore the rate of uplift as observed on the surface will not correspond to the rate of the tectonic uplift, the latter being reduced by the rate of erosion. If the rate of denudation is greater than the rate of tectonic uplift, the earth's surface will be observed to sink.

From the above, it follows that when the accumulation of sediments and erosion are simultaneous, the vertical movements of the earth's crust are not reflected on its surface in their true magnitude. These movements are always distorted, and the distortion is in the direction counter to that of the tectonic movement: because of the accumulation of sediments, the sinking of the earth's surface is slower than the tectonic subsidence, and owing to erosion, its elevation is slower than the tectonic uplift. In some cases, the surface moves in the direction opposite to that of the tectonic movement (elevation of the surface simultaneous with tectonic depression of the earth's crust when the accumulation of sediments exceeds the rate of subsidence, and sinking of the surface simultaneous with tectonic uplift when the erosion exceeds the rate of uplift).

Thus we come to the conclusion that one must differentiate between the tectonic process of uplift and subsidence of the earth's crust, on the one hand, and the manifestation of this process on the surface, where it is distorted by the accumulation and erosion of sediments, on the other hand. This manifestation of the oscillatory crustal movements, expressed as the algebraic sum of the tectonic movements and the surface erosion and deposition, will be called the *geomorphologic expression*, inasmuch as it is reflected in the relief of the earth's surface. Only where accumulation and erosion are absent are the tectonic movements and their geomorphologic expression identical.

RELATIONSHIP BETWEEN OSCILLATORY MOVEMENTS AND THEIR GEOMORPHOLOGIC EXPRESSION

An attempt will be made to explain the relationship that exists between the geotectonic and geomorphologic processes associated with the oscillatory movements of the earth's crust—that is, how to determine the relationships between oscillatory movements and their geomorphologic expression in geologic sections, in the distribution of facies and the thicknesses of the “fixed” sediments that form part of the section.

The “fixed” sediments are understood to be the deposits overlain by the rocks of the latest stratigraphic subdivision (with the exception of Quaternary). The reason for this exception will become clear later.

The attempt to solve this problem will encounter peculiar difficulties typical, as we shall see, of geotectonic problems in general. It appears that the regularities defining the relationship between the geotectonic oscillatory movements and their geomorphologic expression will vary with the scale and the degree of detail in which these movements are studied. These regularities are comparatively simple if the oscillatory movements of the earth's crust are examined in their general aspects only, encompassing large regions and long periods of time and leaving

aside the details of the process. But these regularities become complex as soon as one begins to investigate this process in detail, taking account of its characteristics as they appear in limited areas and in brief periods of time. In the first case, certain average and general regularities are established, whereas in the second, one encounters numerous deviations from these average regularities. These deviations occur in particular cases, but they sometimes cancel each other over the course of time.

Both general and particular regularities are of interest. Knowledge of the first is indispensable in forming the conception of the geotectonic process as a whole and in developing a general theory of this process. The second are necessary in solving local specific geotectonic problems, which are mostly of a practical nature.

At the present level of the science of geotectonics, it has been possible to establish satisfactorily the regularities of a general character and, up to now, to accumulate data regarding particular deviations from these broad regularities. Evidently, the general regularities are only the first approximation in solving the geotectonic problems that confront us. The immediate task of this science must be to develop further approximations by analyzing and generalizing from occurrences that are still in the category of "particular."

The geologic section through the coal measures of the Donbas may be referred to again. It has already been noted that throughout the whole thickness of this section the deposits are exactly the same in nature. The conditions under which they were formed did not change perceptibly during the entire time of their accumulation. In particular, the depth of the basin, at the bottom of which all the sediments were deposited, has not changed. Consequently, during the Middle and Upper Carboniferous the subsidence was in general completely compensated by the accumulation of sediments, so that the geomorphologic expression of the tectonic subsidence was equal to zero.

This conclusion, however, is correct only to the first approximation. It was noted above that the interlayering of beds of coal with other rocks indicates a vertical oscillation of the earth's surface, which either approached sea level or sank again to a slight depth. In the light of what has been said previously, oscillations may be seen as the result either of periodic changes in the rate of accumulation of sediments with a uniform subsidence, or of periodic changes in the rate of subsidence with a uniform accumulation of sediments, or of both these factors. Since, ultimately, there has been no discernible change in the depth of the basin (in the hypsometric level of the earth's surface) and the accumulation of the sediments has sufficiently compensated for the subsidence, the thickness of the accumulated sediments is equal to the depth of the tectonic subsidence of the earth's crust. Consequently, the thickness of the deposits serves as a measure of this subsidence, which

during the Middle and Upper Carboniferous was at least 6.5 km in the territory of the Donbas. Against the background of this figure, the oscillations of the surface that caused changes in the depth of the basin, amounting to a total of a few score meters, were insignificant.

These results may be expressed graphically. In Fig. 85, the abscissa represents the time; the stratigraphic subdivisions are laid off on this axis. The ordinate represents the amplitude of the oscillations. The horizontal line *OP* represents the zero surface, or sea level. Since the oscillations of the surface with an amplitude of few score meters and the movement of the earth's crust with a magnitude on the order of 6 km cannot be depicted on the same scale, oscillations of the earth's surface (the bottom of the basin) coincide with the line *OP*. Below the line *OP* are laid off the thicknesses of the sediments accumulated during the corresponding periods, and the curve *OS* is drawn along the points thus obtained. This last curve represents the tectonic subsidence of the earth's crust; the first represents the geomorphologic expression of this movement on the surface.

As another example, in the vicinity of the city of Ordzhonikidze, on the northern slopes of the Caucasus, the Lower and Middle Jurassic are represented by a series of shales with interbedded sandstones. This series is uniform throughout its thickness of about 3,000 m. The homogeneity of this series testifies that the depth of sedimentation did not change noticeably during the deposition of the series and that the deposition progressed at such a rate as to compensate for the tectonic subsidence of the earth's crust. Overlying these rocks are limestones of the Upper Jurassic, as much as 1,500 m thick; these limestones include coral, oolitic, bituminous, detrital, sandy, and dolomitized varieties. All these are shallow-water limestones deposited at a very small depth (a few score meters). The fluctuation in the depth, it may be assumed, was very small in comparison with the thickness of the limestone series. We may thus conclude that during the Upper Jurassic the bottom of the basin did not undergo oscillation and that the accumulation exactly matched the depression.

It is of interest to determine what took place on the surface of the

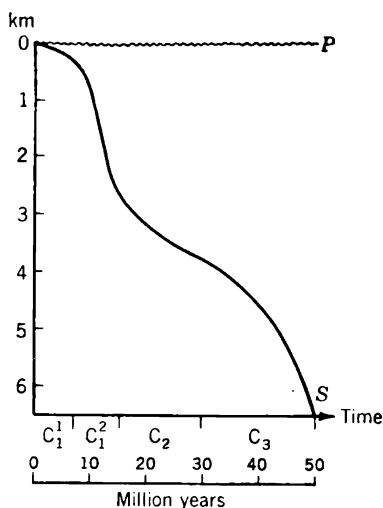


FIG. 85. Diagram of the oscillatory movements in the Donbas during the Carboniferous. *OP* = geomorphologic expression of oscillatory movements; *OS* = geotectonic movement.

earth between the Middle and Late Jurassic. It is hardly possible to establish, with any precision, the depth of the basin during the accumulation of the Lower and Middle Jurassic shale series. But the presence of sandstone layers in these series and of coal seams and detrital limestones in the neighboring regions strongly suggests that this depth was very small. Possibly the average depth was somewhat greater than the depth of deposition of the Upper Jurassic limestones and reached 100 to 200 m. If this was the case, toward the Late Jurassic the depth of the sea diminished and the earth's surface was somewhat elevated. Thus it is surmised that between the Middle and Late Jurassic the rate of sedimentary accumulation increased, so the thickness of the shales and limestones is greater than would have been required to compensate for the tectonic subsidence, the excess thus representing an "overload" of sediments, which at this time were accumulating more rapidly. How great was this "overload"? The figures cited above show that it cannot have exceeded 100 to 150 m. Thus, of a total thickness of Jurassic deposits equal to 4,500 m, about 97 per cent (4,350 m) were formed by compensation of the depression through the accumulation of sediments, and only three per cent is the excess of deposition over subsidence. It may also be surmised that between the Middle and Late Jurassic there was uplift of the earth's crust, causing the basin to become more shallow; thereafter subsidence again occurred, but it was continually compensated by the accumulation of sediments.

The Lower Cretaceous deposits of the same part of the Caucasus are composed of interbedded marly and shaly sandstones, sandy clays, and, to less degree, of marls and organogenic limestones. The characteristic mineral is glauconite. The depth of the basin remained very small, hardly more than 100 m. One may therefore assume that the bottom sank very little (a few score meters) after the Late Jurassic. The thickness of the Lower Cretaceous, however, reaches 1,150 m.

The Upper Cretaceous limestones consist almost entirely of Foraminifera and of fragments of the prismatic layer of *Inoceramus* shells. These sediments were earlier believed to be of the deep-water type, but the presence of the fragments of coarse shells contradicts this supposition. The basin may have become somewhat deeper, but no deeper than 200 m. The thickness of the Upper Cretaceous is about 300 m. The sinking of the bottom resulted from the fact that the movement of the earth's crust was not compensated by the accumulation of sediments. This absence of compensation may have been due either to slowing of the process of accumulation or to a greater rate of subsidence of the earth's crust relative to that of accumulation. Here the relationship between the geomorphologic expression and the tectonic movement is not the same as in the previous case: the bottom may have been lowered by 150 m at the beginning of the Late Cretaceous, and the thickness of

sediments of this time is 300 m. The tectonic subsidence is therefore 450 m.

In the preceding cases, when the thicknesses were great and the oscillations of the surface were small, the magnitude of the crustal movement could be determined from the thickness of the sediments, disregarding a certain incompleteness in the compensation of the subsidence by sedimentation, since this discrepancy was very slight. By following the same procedure in the last case, the error would have been about 30 per cent. If the thicknesses of the entire Cretaceous system are added together, in comparison with this total the change in the hypsometric level of the bottom of the basin between the Early and Late Cretaceous will again be negligible (10 per cent).

In the foothills of the Caucasus a section of Tertiary deposits may be studied. On the northern slope these sediments begin with a group of marls belonging to the Paleocene, Eocene, and Lower Oligocene, the total thickness being 170 m. They are overlain by the Maykop suite, which is 850 m thick and consists almost entirely of clays. Above this there are Miocene deposits, which (the lower beds of the Upper Sarmatian stage) are represented by marine shallow-water, sandy, and shaly sediments of tremendous thickness (2,000 m). The upper beds of the Upper Sarmatian stage, the Meotis, and the whole of the Pliocene (800 m) are composed of continental sediments, mainly conglomerates, whose coarseness increases from the bottom upward.

A survey of this section indicates that the surfaces of the lithosphere in the northern Caucasus rose during the Tertiary period. The Eocene marls were probably deposited at a greater depth than the Maykopian clays; in any case, the Miocene sediments appear to be of a more shallow-water type than the Paleogene deposits. At the end of the Miocene the deposition surface rose above sea level. What was the magnitude of this uplift? The depth of formation of the Paleogene beds was no greater than the depth of deposition of the Upper Cretaceous, which did not exceed 200 m. The continental series of the Pliocene were evidently formed on the low plains of the littoral zone, whose surface was certainly no more than 100 m above sea level. Thus the surface of the lithosphere was raised as much as 300 m; the thickness of the Tertiary sediments, however, reached 4,000 m. It follows that only 7 to 8 per cent of the whole thickness of the series represents the "overload" of sediments and the rise in their surface. The remaining 92 to 93 per cent is the result of compensation for the subsidence of the earth's crust by the accumulation of sediments. This case is an example of elevation of the surface of the earth's crust superimposed on a background of intensive tectonic subsidence, whose magnitude therefore is

$$4,000 - 300 = 3,700 \text{ m}$$

The results of this analysis are shown in Fig. 86. This diagram reveals at a glance the small amplitude of the oscillations of the earth's surface in comparison with the amplitude of the tectonic movements and how closely the accumulation of sediments compensates for the tectonic subsidence.

It is interesting to determine the course of the same processes in regions where the sedimentary series have comparatively small thicknesses, as, for example, on the Russian plain. Roughly estimated, the accumulation of sediments in the Volga region during the Middle Devonian, through and including the Kazanian age, exceeds 2,000 m. The depth of the basin in Kazanian time differed little from that in the Devonian. Here, as before, the oscillations of the surface during an extended period were extremely small in comparison with the magnitude of thicknesses.

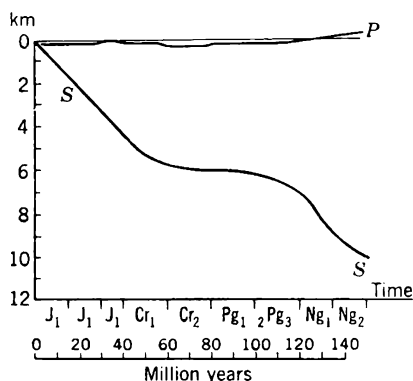


FIG. 86. Diagram of oscillatory movements in the vicinity of the city of Ordzhonikidze. Symbols are the same as those in Fig. 85.

previous cases, but it is still not great: the earth's crust here underwent subsidence of 650 m, of which 500 m was compensated by accumulation of deposits, so that only the remaining 150 m are the geomorphologic expression of the tectonic movement.

When the gypsum-dolomite and limestone rocks of the Kazanian stage are replaced by the multicolored deposits of the Tatarian stage, a certain part of the latter's thickness is undoubtedly the result of a general elevation of the surface of the sediments. But this portion is not great: the Kazanian deposits in the Volga region have a thickness of about 150 to 170 m, and the Tatarian deposits are about equally thick. Of this total thickness of 300 m, the elevation of the earth's surface constitutes no more than 100 m, since the depth of the Kazanian basin was very small and the Tatarian deposits were formed on the low-lying flats of the littoral zone. Here about 30 per cent of the total thickness is

An examination of shorter spans of time naturally yields somewhat different relationships. Thus, for example, from the beginning of the Middle Jurassic to the end of the Upper Cretaceous about 500 m of deposits were accumulated in the Volga Basin. The Middle Jurassic begins with very-shallow-water sediments, and the Upper Cretaceous is represented by marls and chalk. The depth of the basin may have increased by 100 or 150 m during this time. Here the relative change in the elevation of the surface of the sediments is greater than in the previous cases, but it is still not great: the earth's crust here underwent

"overload," and the remaining 70 per cent compensates the subsidence. Thus it appears that in regions of relatively thin strata the geomorphologic expression of the tectonic movement is also considerably smaller than the movement, and the magnitude of the sedimentary accumulation differs little from that of the crustal subsidence.

Even without recourse to the examples cited above or any others, the conclusion that the amplitude of the geomorphologic oscillations is, as a rule, small in comparison with the tectonic oscillations, in the area of subsidence, is perfectly valid. It must be remembered that geologic sections for the most part are composed of either shallow-water (neritic) marine sediments deposited at depths not exceeding 200 m or continental deposits formed on low-lying alluvial plains. It was believed not long ago that among the sediments observed in geologic sections

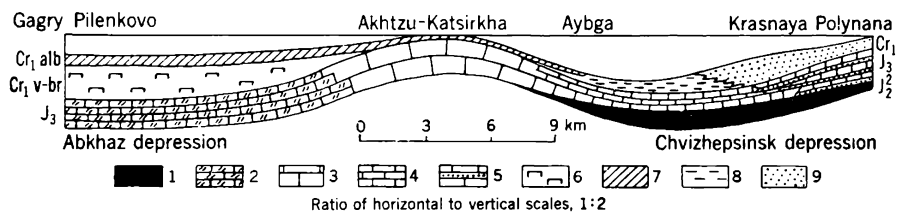


FIG. 87. Stratigraphic section through the Jurassic and Lower Cretaceous deposits along the Gagry-Krasnaya Polyana line. Middle Jurassic deposits: (1) shales and sandstones of the Al'ginskaya suite of the Bathonian stage. Upper Jurassic deposits: (2) asphaltic and lithographic limestones (breccial limestones at the top); (3) massive, nonbedded limestones; (4) platy limestones and marls (basal and conglomeratic limestones); (5) flysch series of the Tithonian stage. Lower Cretaceous deposits: (6) bedded limestones (limestones and marls at the top); (7) dark Albian marls; (8) Lower Cretaceous marl series; (9) Lower Cretaceous flysch series (after B. M. Keller).

there are also deep-water deposits (bathyal or even abyssal). These, however, have recently been reexamined: the new conclusions are to the effect that shallow-water deposits predominate. Thus the chalky marls, siliceous shales, black bituminous shales, and many other sediments are considered by present researchers to be of shallow- rather than deep-water type.

The clear predominance of shallow-water sediments in geologic sections indicates that, as a rule, the subsidence of the earth's crust has been compensated by the accumulation of sediments. This question has been specially studied by B. M. Keller (1948), in order to discover cases in which the compensation has been interrupted. Such cases have actually been found. One of these relates to the Gagry-Krasnaya Polyana section in the Caucasus, which includes two zones of great subsidence of the earth's crust (the Abkhaz and Chvizhepsinsk troughs) and a zone of less subsidence between them (the Akhtsu relative uplift).

As one may see from the section (Fig. 87), the Jurassic and Cre-

taceous strata as a rule have a greater thickness in the depressions and a smaller thickness (down to complete wedging out) over the uplift. Thus there is compensation of the crustal movements by the accumulation of sediments in this case as well. This relationship, however, is abruptly disturbed during the Late Jurassic. In the Chvizhepsinsk subsidence the Upper Jurassic sediments are represented by a series of platy limestones and marls up to 300 to 400 m in thickness. In the Akhtzu uplift the Upper Jurassic deposits are represented by massive reef limestones; here their thickness, instead of being smaller, is greater than in the depression and reaches 600 to 700 m.

B. M. Keller explained this unexpected increase in the thickness of the sediments in the Akhtzu area by a temporary retardation in the accumulation of sediments in the Chvizhepsinsk subsidence. There was no change in the relationship between the movements of the crust in the

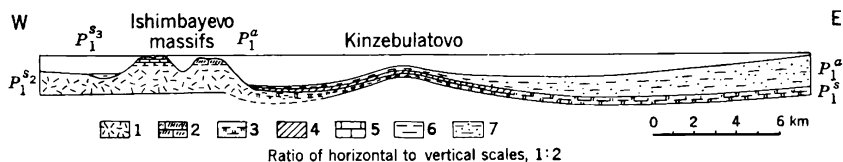


FIG. 88. Stratigraphic section through the Sakmarian and Artinskian deposits in the Ishimbayevo area of the Cisurals. Sakmarian deposits: (1) massive reef limestones of the Tingskii and Tastubskii strata; (2) massive reef limestones of the Sterlitamakskii stratum; (3) dolomites, marls, and argillites; (4) the same rocks, belonging to the Sterlitamakskii stratum. Artinskian deposits: (5) massive reef limestones; (6) marls and argillites; (7) marls and argillites with interbedded sandstones (after B. M. Keller).

two neighboring regions even for a short span of time. This retardation is associated with the facies of the sediments. During this time, over the more slowly subsiding Akhtzu uplift, reef limestones were rapidly forming and thus fully compensating the subsidence of the earth's crust. In the Chvizhepsinsk depression, where the subsidence was progressing faster than the accumulation of platy limestones and marls, the compensation was falling short of the subsidence. During the Cretaceous period, the earlier normal relationship was reestablished.

The second example relates to the Ishimbayevo Cisural region, east of the well-known reef massif of Artinskian age (Fig. 88). The sediments developed in the structural depression east of and contemporary with these reef limestones differ considerably from the latter, having a much smaller thickness. These deposits are represented by a series of clays, marls, and dolomites up to 200 m thick, whereas the thickness of the reef facies exceeds 1,000 m. In this case, the subsidence was evidently uniform throughout the whole region. On the reef mass this subsidence was compensated by the accumulation of limestones, but east

of it the accumulation of sediments lagged considerably behind the subsidence of the earth's crust. Such an equalization took place in the Chvizhepsinsk basin during the Cretaceous and Paleogene and in the Ishimbayevo basin during the Kungurian age.

A supplementary illustration is provided in Fig. 89, which shows

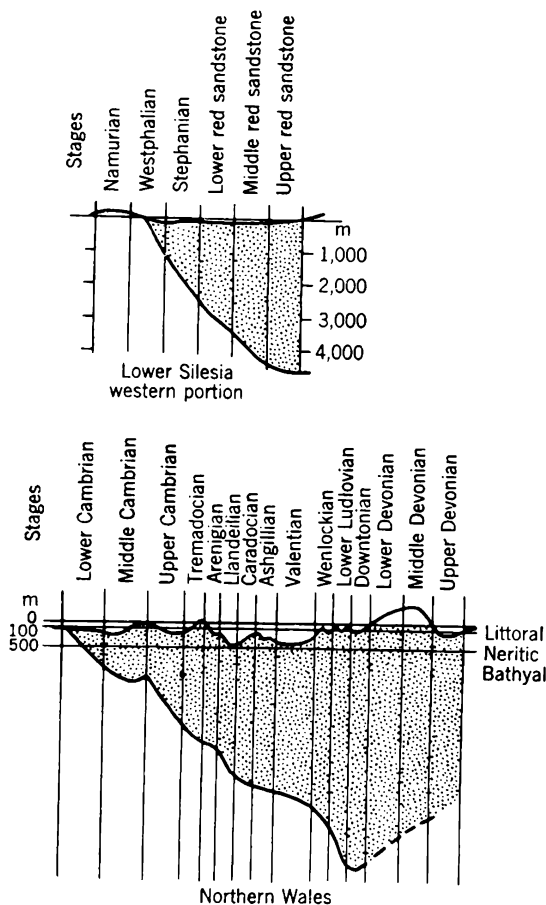


FIG. 89. Curves of oscillatory movements. On each graph, the upper curve is the surface movement and the lower curve is the tectonic movement (after S. Bubnoff).

a few "epeirogenic curves" of different regions of Europe. These curves were constructed by S. V. Bubnoff (1935) after the manner used earlier in this book. It can be seen conclusively at a glance that the amplitude of the surface oscillations is small in comparison with the tectonic subsidence.

Thus we come to the conclusion that if there are deviations from the normal compensation of subsidence by the accumulation of sediments

in the regions of crustal subsidence, these deviations are as a rule very slight. To the first approximation, it may be said that the magnitudes of accumulation and of subsidence are about equal.

Regions in present-day environments in which subsidence is being compensated by accumulation of sediments must be sought in shallow-water basins with smooth bottoms. Shelves are an instance of such areas. Within the shelf are level areas whose depth fluctuates very little and usually does not exceed 200 m. The gently sloping littoral-zone alluvial plains are another example of such regions with a smooth surface. A shelf is formed during the subsidence of the earth's crust. The accumulation of thick sedimentary strata on littoral-zone plains also attests to the subsidence of the earth's crust. This crustal subsidence is, of course, irregular, proceeding at different rates in different places, so the "tectonic relief" of the region must be complicated. The fact that the surface of the deposits is smooth both on the shelves and on the littoral plain indicates that at different points a different magnitude of crustal subsidence is compensated by a correspondingly different thickness of sediments. As a result, the oscillatory movements of the earth's crust are not reflected on its surface.

This question naturally arises in connection with deep-water marine basins and oceans. These contain the deepest depressions in the relief of the lithosphere that are not isostatically compensated (and this is undoubtedly the case). The relationship between the tectonic movement and their geomorphologic expression is quite different here. The Philippine trench, which in places is more than 10 km deep, was formed by subsidence of the earth's crust. Because of the almost complete lack of any accumulation of sediments, the geomorphologic expression of the tectonic movement coincides, in this case, with the movement, so that the "tectonic relief" becomes the same as the surface relief.

The tectonic significance of phenomena that occur in the oceans will be considered in detail later. Here it may be said only that the geologic sections through the present continents contain no indications of deep-water marine deposits; there is no factual basis for asserting that the ocean basins ever existed on the sites of the present continents. Inasmuch as there are no signs of oceanic conditions prevailing in sections available for study, the lack of compensation of subsidence by accumulation in the present oceans is so far of no importance, since it is not observed in objects that may be studied by a geologist on dry land.

We may now turn to regions of uplift. Has any general compensation of the tectonic uplift by erosion in the regions of uplift of the earth's crust been observed, analogous to the compensation of subsidence by accumulation? At first glance it appears that the answer should be negative. High mountains and a relief intricately carved by erosion are encountered in regions of uplift. It appears that here the geomorphologic

expression of the oscillatory movements is much more complicated and cannot be reduced to a simple formula of compensation by erosion.

Let us, however, scrutinize this problem more deeply. The relief that is now observed on the surface of continents is young. Investigation of ancient surfaces of erosion in mountainous regions leads geomorphologists to the conclusion that mountain ranges, as they are seen at present, have appeared quite recently in the geologic scale of time. The Caucasus range, for instance, was peneplaned and uplifted again several times in the Late Tertiary and Quaternary. Similarly, the relief of the Urals, where the uplifted Quaternary erosional surfaces indicate that not long ago there was a peneplain on the site of the present range, is very young. The relief of the continents is in itself transient and rapidly changing. Observation of the degree of preservation of the former relief of the earth's surface in geologic sections does more than anything else to convince one of the validity of the above statement. It was said earlier, in discussing the compensation of crustal subsidence by the accumulation of sediments, that only the strata "fixed" in the sections, the covered "root" deposits, are of concern. The same applies to regions of uplift: in establishing the general, average regularities one must determine how the region of uplift is reflected in "fixed" geologic sections. By following this procedure, it will readily be seen that the ancient relief of former geologic periods is not preserved in geologic sections, as a rule. For example, mountains undoubtedly existed at the end of Paleozoic in the area of the present Donets basin. The Mesozoic sediments, however, lie on an eroded folded Paleozoic basement and thus on an entirely horizontal surface, if one disregards dislocations that took place in a much later period. Thus the ancient relief was completely destroyed, and the surface of the region was made horizontal. The geologic section contains no buried mountains and no valleys covered with later sediments.

Like the Donbas, shortly after their formation, the Hercynian mountains of western Europe, western Siberia, and many other regions were completely reduced to a horizontal surface. At the end of Precambrian times there were undoubtedly many high mountain ranges on the earth; toward the beginning of the Cambrian, however, they were all leveled, and the Cambrian sediments were deposited on a horizontal surface.

A few random examples have been cited. But the great majority of observed instances of deposition of beds with angular unconformities may also be grouped in the same category with these examples, since they indicate that the surface upon which these unconformable overlying beds were deposited was usually horizontal.

On platforms, where the beds were not contorted into folds, where no mountains were built and where only the usual relief of a plain with shallow valleys and low hills could have existed, one also finds no

remnants of such a relief on the surfaces of unconformities. As a whole, there is a great predominance of smooth, primarily horizontal planes of contact between suites, even those separated by brief gaps of time.

The destruction of the terrestrial relief within the compass of geologic time is again the most common regularity. In individual instances deviations from this regularity are observed, but they are temporary and usually very small, just as in the case of compensation of subsidence by accumulation. Thus at the top of the Carboniferous limestones in the Moscow basin there are fairly large irregularities, such as buried valleys up to 50 m in depth. In certain regions of Kazakhstan a hilly Paleozoic relief is buried under the Mesozoic sediments. In Uryupin rayon of the Stalingrad oblast' the sediments of Khoper stratum of the Danian stage overlie Turonian white chinks greatly dissected by ancient erosion. The depth of these erosional valleys reaches 20 to 25 m (Keller, 1948). A number of erosional depressions have been noted in particular areas of Paleozoic sediments in the Volga region. At Kama, for instance, on the so-called Buldyr uplift, which is a tectonic dome, in a small depression between two separate peaks the lower stratum of the Lower Carboniferous appears to be eroded, and the overlying coal measures fill up this erosional depression. On the Syzran uplift at Samar Luka the same coal measures lie upon the eroded surface of the Chernyshinsk limestones, the top surface of which contains deep erosional depressions. Again at Samar Luka, in the region of Kostychenko uplift, the Jurassic sediments lie on a greatly eroded surface of Paleozoic rocks (Rozanov, 1949).

Ancient erosional irregularities are better preserved in the buried state where they are overlain by strata of continental origin. The marine abrasion associated with the transgression of marine deposits almost always obliterates every irregularity in the surface. Wherever a buried erosional relief exists, this relief affects the distribution of the thicknesses: the sediments filling the depressions in the relief are of greater thickness than the contemporary sediments covering the projections in the buried relief. This variation in the thickness apparently has no relation to the oscillatory movements of the earth's crust.

All similar cases belong to the category of exceptions, of which it has been said that "the exceptions prove the rule." Although buried relief does exist, its scale in general is very small in comparison with that of the relief now existing on the surface of the earth. Buried relief is not commonly encountered, and as the sediments under which it is buried accumulate, the relief rapidly flattens out upward in the section: within a few score meters upward the relief is usually not manifested either in the occurrence of beds or in the distribution of their thicknesses. Thus, keeping in mind the general rule, it may be said that the dissected relief of regions of uplift and erosion is not preserved in the geologic

section: the relief appears to be almost entirely destroyed before the deposition of the succeeding beds, and the exceptions to this are obviously of only secondary importance.

It should be remembered that this discussion concerns the occurrence in the geologic section of pre-Quaternary rocks that are also covered by pre-Quaternary rocks. The conditions of deposition of Quaternary sediments are entirely different: these fill depressions, lie on buried slopes, and one can indeed find buried relief under these sediments almost anywhere. The reason for such differences between the more recent and more ancient sediments will become clear later. It will be seen that the reasons lie not in the peculiar character of this epoch but in the greater interval of time during which the pre-Quaternary sediments were eroded, redeposited, and redistributed. Occasionally it may have happened that the time interval was not sufficient in the case of young Tertiary deposits as well, and in these one may also find inclined primary bedding and other forms of adjustment to the relief. Such cases are observed very seldom in more ancient sediments. On the other hand, even in the Lower Quaternary sediments, the bedding may be horizontal, their thickness may compensate for the subsidence of the earth's crust, and they may have other characteristics of pre-Quaternary formations. This is observed in zones of recent and intensive subsidence, such as the Terak and Kura depressions of the Caucasus.

A conclusion may be formulated as follows: keeping in mind the most general features of the processes of accumulation and erosion in those parts of the earth's surface that are accessible to geologic investigation (that is, the continents), it may be said that the deposition and erosion destroy both the positive and the negative features of the relief that has been produced by oscillatory movements and surface erosion. Subsidence and depressions are compensated by accumulation and uplift and projections by erosion. Scarcely any traces of ancient relief are preserved in the geologic section. In other words, over any extended period of geologic time the geomorphologic expression of oscillatory movements is negligible. In the structure of the crust, however, the relief is broadly imprinted in a different manner—in the facies of deposits and their distribution, which will be dealt with later.

If one is concerned not with broad general regularities but with particularities manifested in local and short-lived phenomena, one cannot disregard the deviations from "the middle line," which were discussed above. In this case, one must consider both the deficiency in the compensation of subsidence by accumulation, which locally can exist for a certain period of time, and the possibility that there is a buried relief of no great magnitude. These particularities acquire great importance in detailed investigations of problems of an applied nature (study of the structure of a specific oil, bauxite, or iron deposit or of a particular part

of a coal basin, engineering geology studies, etc.). At the present state of development of geotectonics, however, the deviations from the average rule that may be noted are not generalized, and the laws governing such "exceptions" have not yet been established, if they exist. The search for regularities in the appearance of "exceptions" is one of the most pressing and, from the utilitarian point of view, most important tasks of this branch of geotectonics.

CHAPTER 13

FACIES AND THICKNESS OF FORMATIONS IN RELATION TO OSCILLATORY CRUSTAL MOVEMENTS

CORRELATION OF FACIES AND THICKNESS OF SEDIMENTS

The preceding chapters indicate a relationship between oscillatory crustal movements and the thicknesses and facies of sedimentary formations. This chapter will investigate in more detail the causes of this relationship and its manifestations.

We have established that subsidence of the earth's crust is approximately compensated by accumulation of sediments, and uplift by erosion. Subsidence may be considered a result of the pressure to which the earth's crust is subjected under the weight of the deposited sediments, and uplift the result of the relief of pressure on the crust during erosion. Such hypotheses have been frequently suggested. They are based on a supposition that the earth's crust is floating on a viscous substratum and that during additional loading the crust is slightly immersed in this substratum; with a diminution of its weight, it emerges.

To decide whether this method can explain an observed correspondence between subsidence and accumulation or between uplift and erosion, one must first establish a relationship between thickness and lithologic facies. This is dictated by the following reasoning: if subsidence of the earth's crust is induced by the weight of sediments, then the rapidity of subsidence depends on the rate of accumulation. And the distribution of zones of more rapid and slower accumulation of sediments should inevitably change with the change in distribution of lithologic facies.

Let us imagine that, owing to delivery of abundant detrital material from a nearby, rapidly uplifted source area in a certain zone, great thicknesses of terrigenous sediments are rapidly accumulated. These thicknesses decrease away from the source of material. Therefore, for a certain time, a certain distribution of greater and smaller thicknesses exists: in the hypothetical zone described they are greater; beyond it

they are smaller. Let us assume that after a certain time the environment is changed; the erosion of the land ceases, or diminishes, and in the basin sediments of another facies, such as limestone, begin to be deposited. The rate of accumulation of the limestone is determined by factors other than those which control the rate of accumulation of clastic sediments. The formation of limestone depends on the activity of organisms, and this activity in a marine basin depends on the temperature and salinity of the water, its transparency, currents, depth, and other such factors. A favorable combination of all factors is necessary for the rapid accumulation of limestone. It is difficult to imagine that such a favorable combination is created in the same zone in which a rapid accumulation of clastic material was previously determined by entirely different factors. Even if this happened as an accident, it could not be repeated frequently. More probably, the environment necessary for rapid deposition of limestone will not arise where favorable conditions previously existed for the accumulation of clastic material. Therefore the distribution of the zones of greater and smaller accumulation will change with the change in facies, and if subsidence of the crust is due to the weight of sediments, then the zones which will subside most rapidly are not those which subsided intensely before (Beloussov, 1938-1940; 1940).

The oftener the facies change, the less likely a chance coincidence of zones of greater and smaller accumulation of deposits in the change in facies. In a complex section containing alternating rocks of different composition and origin, we should be able to observe constant changes in the disposition of the zones of greater and less accumulation, that is, the zones of greater and smaller thickness. But in switching from theoretical reasoning to the study of real geologic sections, we find that the expected direct relationship between the distribution of zones of greater and less accumulation and that of the lithologic facies is not observed.

If we compare the distribution of facies and thicknesses of deposits in the European part of the U.S.S.R., we will see that the distribution of thicknesses of sediments—the locations of areas of greater and less accumulation—is considerably more stable in time than the distribution of facies.

During a prolonged geologic time, sediments of greater thickness accumulate in a given area, regardless of the change in facies or sedimentary environments. The correlative adjacent zones show an accumulation of smaller thicknesses, also independently of the facies.

Figures 90 to 94 show the distribution of thickness and facies of the deposits on the Russian plain for several stratigraphic units. This distribution is plotted by isopachs: the points at which the thickness of the sediments of a given age is the same are connected by isopachs (Ronov, 1949).

After the second half of the Devonian period, when general subsidence turned the Russian plain into a basin of deposition, the thicknesses of the sediments whose lithology and distribution are represented in Fig. 90 were established. The areas of relative thickness occur in a meridi-

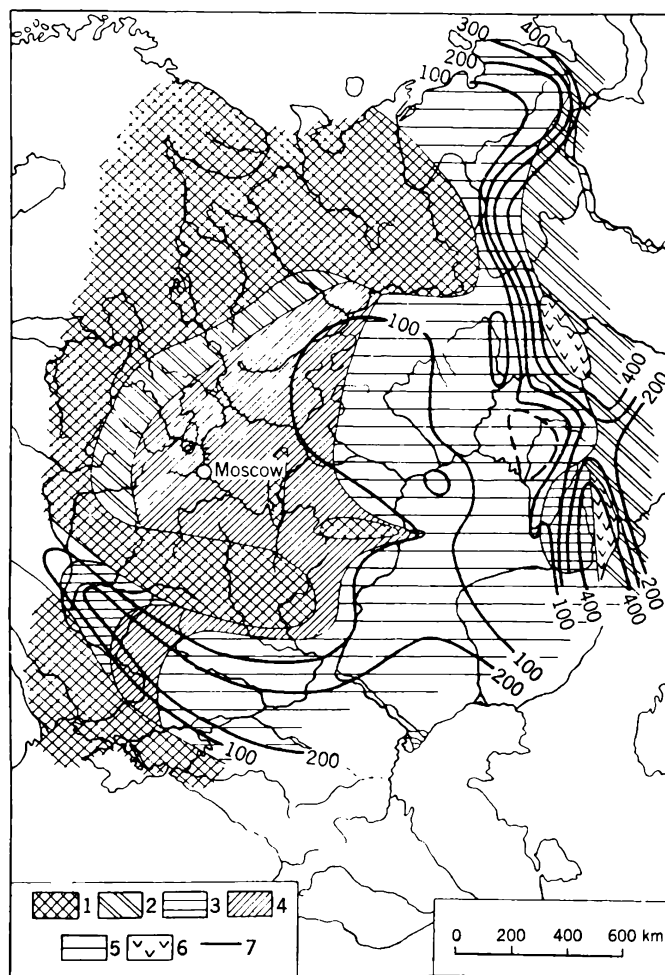


FIG. 90. Facies and thicknesses of Tournaisian deposits on the Russian plain: (1) zones of erosion; (2) coal-bearing deposits; (3) sandstone and shale; (4) limestone, shale, and sandstone; (5) limestone and dolomite; (6) volcanic rocks; (7) isopachs in meters (after A. B. Ronov).

onal zone extending from the Caspian Sea to the lower part of the Pechora River. From this zone a "bay" of areas of great thicknesses of deposits turns westward through the Moscow basin. West and south of the Moscow basin, the thickness of the sediments decreases toward the Baltic and the White Sea and also toward Voronezh, where an eastward

“cape” of smaller thickness is observed. Farther south the thickness increases. The other details of the distribution of thicknesses are easily understood from the map.

The distribution of the zones of greater and smaller accumulation

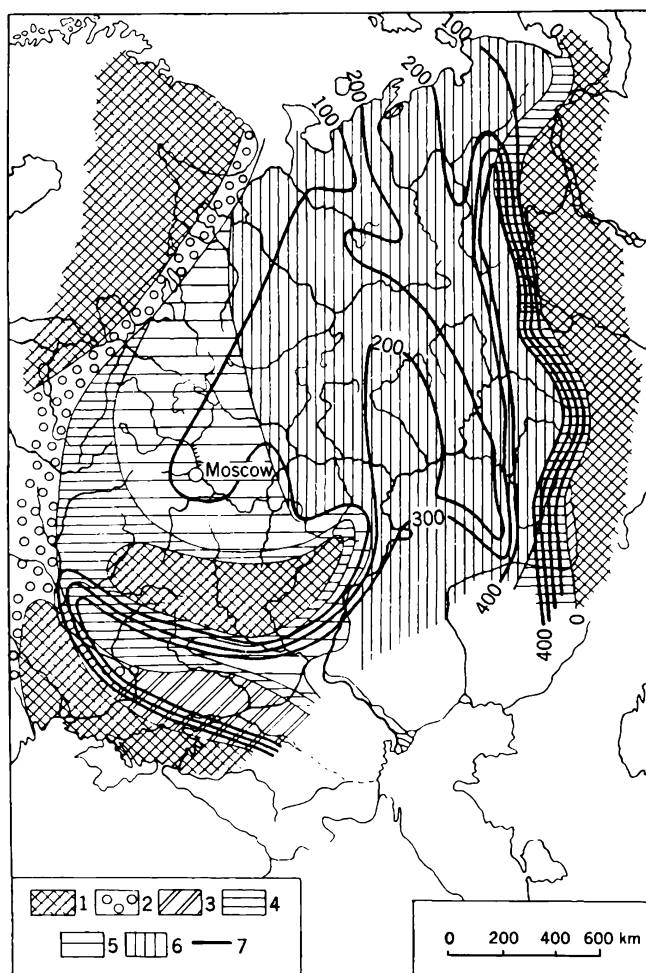


FIG. 91. Facies and thicknesses of the Upper Carboniferous on the Russian plain: (1) zones of erosion; (2) red beds; (3) coal-bearing formations; (4) sandstone and shale; (5) limestone interbedded with shale; (6) limestone and dolomite; (7) isopachs in meters (after A. B. Ronov).

was extremely persistent and was maintained through the Cretaceous. During a very long period of geologic time, sediments of relatively great thickness were accumulated in one place, while at the same time in other places sediments of less thickness were deposited with the same persistence. Some changes are observed, but they do not alter the basic

picture: in one case, the region of small thicknesses expands at the expense of the adjacent zones of great thickness; in another, the opposite changes occur. The quantitative expression of the differences in thickness also changes; in certain cases the difference between the thick-

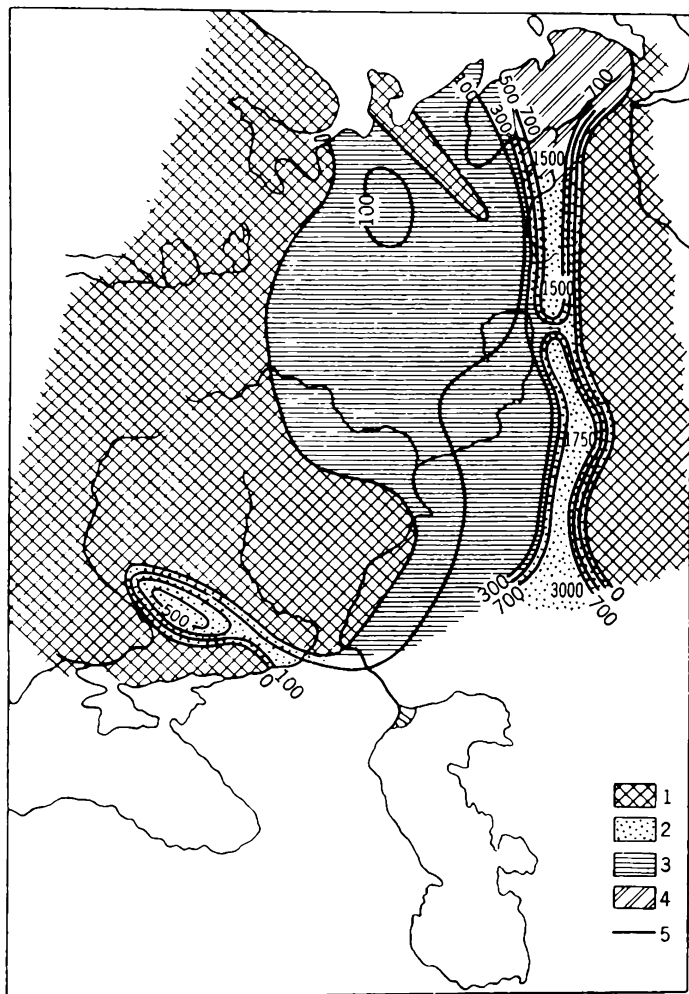


FIG. 92. Facies and thicknesses of the Artinskian stage of the Lower Permian on the Russian plain: (1) zones of erosion; (2) sandstone and shale; (3) coal-bearing deposits; (4) limestone and dolomite; (5) isopachs in meters (after V. V. Belousov and A. B. Ronov).

nesses in the zones of greater and smaller accumulation is reduced to several tens of meters; in others, it reaches hundreds of meters. But during all these changes the general plan is preserved, as reflected in all maps by the characteristic bending of the isopachs contouring the Volga region, the Moscow basin, the Dnepr-Donets region, etc.

The stability of the distribution of thicknesses contrasts with the changeability of the facies distribution. From the Middle Paleozoic to the end of the Late Cretaceous on the Russian plain, limestones of different types, marl, chalk, sandstone, various shales, lagoonal, gypsum-

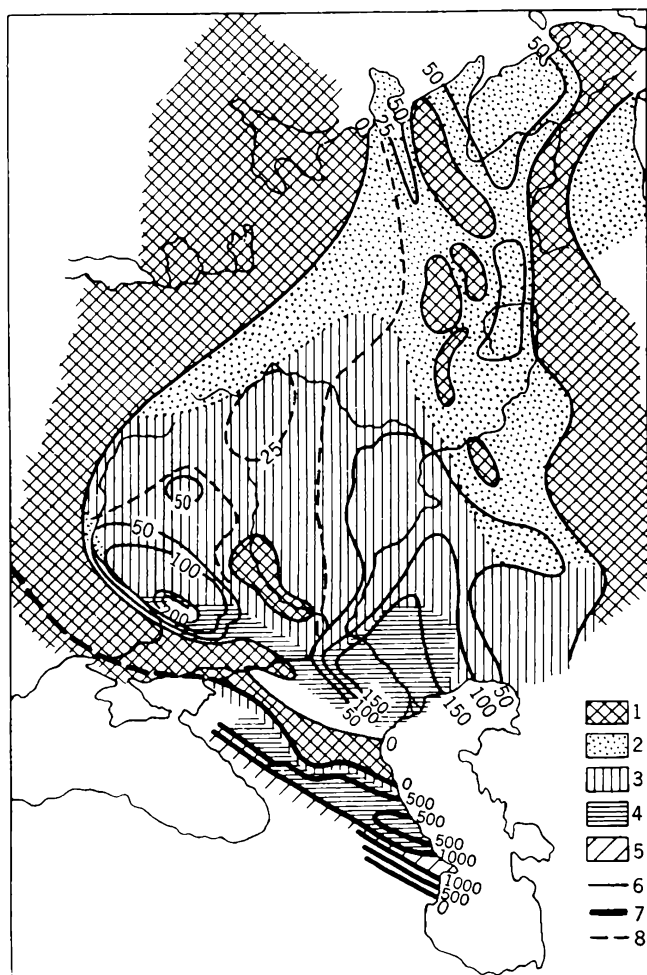


FIG. 93. Facies of deposits and thicknesses of Upper Jurassic sedimentary rocks on the Russian plain: (1) zones of erosion; (2) sandstone and shale; (3) shale; (4) limestone and marl; (5) calcareous flysch; (6) isopach contour interval, 50 m; (7) isopach contour interval, 500 m; (8) isopach interval, 25 m (after V. V. Beloussov and A. B. Ronov).

and salt-bearing sediments, and continental red beds were deposited. The thicknesses of these different rock types were distributed, however, in accordance with a certain plain: where relatively large thicknesses of sandstones and shales accumulated at one time, relatively large

thicknesses of limestone accumulated at a later time, and still later relatively large thicknesses of red beds, etc., were deposited, all in the same region.

In a previous work (Belousov, 1940) we showed that in the main

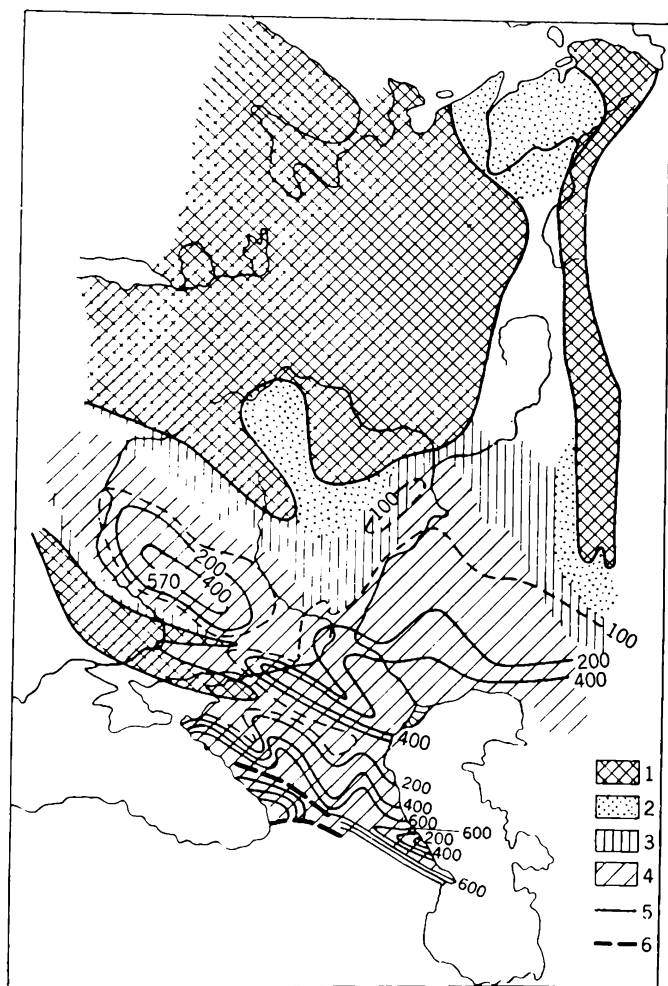


FIG. 94. Facies of Santonian substage and thicknesses of Upper Cretaceous rocks on the Russian plain: (1) zones of erosion; (2) sandstone; (3) shale, marl, sandstone, and diatoms; (4) marl limestone; (5) isopachs at 200-m intervals; (6) isopachs at 100-m intervals.

Caucasus range during the Jurassic, the Cretaceous, and the Paleogene the distribution of the zones of greater and smaller accumulation of sediments was the same, although the lithologic facies changed many times. The zone of maximum thickness has always extended along the southern slopes of the present-day main Caucasus range, including the

zone of flysch facies. Another zone of relatively large thicknesses of deposits occurred along the present-day foothills of the northern slopes. A zone of small thicknesses was situated between these two. In spite of this constant distribution of thicknesses, the facies are highly changeable: the Lower and Middle Jurassic beds are shale and sandstone; the Upper Jurassic are limestone, gypsum, red beds, and dolomite; the Lower Cretaceous are predominantly terrigenous and the Upper Cretaceous beds are calcareous and marly.

Many similar examples may be cited, clearly showing that the distribution of zones of greater and less accumulation is apparently independent of the lithologic facies. It is a manifestation of a more general phenomenon than the arrangements of facies. Consequently, the hypothesis that greater subsidence of the earth's crust is due to more rapid accumulation of sediments is probably incorrect. During the Jurassic and Lower Cretaceous, a greater quantity of thick terrigenous deposits were accumulated in the Volga region than elsewhere. This thick wedge of sediments can be explained by the direction of the streams bringing the clastic material in great quantity to this area. In the same zone, however, both earlier (in Late Devonian and Carboniferous time) and later (in Late Cretaceous) limestones were also deposited in greater thickness than elsewhere. But the more rapid accumulation of the carbonates is determined, as already noted, by entirely different geomorphologic factors. It would be impossible for such very different conditions and forces of nature to act in such harmony that greater and smaller accumulations of sediments always occurred in areas of certain lithologic facies and that these relationships were always maintained over prolonged periods of time. The stability of the distribution of zones of greater and less accumulation, independent of the facies, testifies that the thickness of the sediments is determined by a more general factor acting uniformly on the sediments of any facies.

FACTORS CONTROLLING THE THICKNESSES OF DEPOSITS

This conclusion regarding the independence of the thicknesses and the lithologic facies persuades us that the thicknesses are not determined by the same geomorphologic factors that determine the facies of the sediments. The differences in the thicknesses of deposits must be explained by other processes.

This conclusion, however, seems to contradict observations of the accumulation of recent sediments both in the ocean basins and on the surface of the continents. Actually, in recent times the rate of accumulation in marine basins is very different from that of the geologic past. Great masses of clastic material are rapidly piling up near the mouths of rivers. But the low and shallow shores are now yielding little terrig-

enous material. Corals are rapidly building their structures, and reef limestones are accumulating more rapidly than limestones of any other type. Briefly speaking, the rate of accumulation of sediments under present conditions undoubtedly depends on the same factors that determine the composition of these sediments: on the nature of the shore, the ocean currents, the salinity of the water, its temperature, etc. Because of the diversity of these factors, the thickness of sedimentary bedding at the bottom of marine basins varies widely and depends largely on the character of the sediments available.

The relationship between the type of sediment and its thickness is also revealed in recent continental deposits, where the thickness of sedimentary beds depends on the surface configuration in the area of sedimentation; this, in turn, determines the character of the sediments. The accumulation of pebbles and sand is observed in the estuaries of rivers; other sediments are deposited in lakes; coarse rock waste is piled up at the foot of steep slopes, etc.

Thus a strange discrepancy arises between geologic sections of the remote past and the present. Does the present differ sharply in this respect from all preceding geologic time, or must one look in another direction for the reasons for this apparent contradiction?

In solving this problem, we should bear in mind the process of formation of sequences of thick sediments already fixed in the geologic section, that is, covered by more recent deposits. Our conclusions will not be correct for unburied deposits. This reservation is especially important.

Let us consider the general conditions of sedimentation on the bottom of a marine basin (Zenkovich, 1940; Belousov, 1942, Makowski, 1937).

The clastic material that is a result of disintegration of the land, on reaching the bottom of a basin, is affected by the movements of water causing its displacement along the bottom. Investigators have often supposed that large circular marine currents play the major part in this displacement. In reality the omnipresent vibration of the water particles is of considerably greater significance.

During undulatory movement of the sea, the particles of water move in approximately circular orbits: upward (in the direction of the wind), downward, and backward. With increasing depth, these vibrations become weaker and finally die out. The depth of penetration of the wave of vibrations depends on the force and duration of the wind. In the shelf zone of the open sea it reaches a maximum of 200 m. This depth is called the *base of wave action*. All sedimentary material situated above the base of wave action is subjected to ceaseless washing during the undulatory vibrations of water associated with wave action and is constantly transported from place to place. Stable accumulation of sediments is possible only below the base.

As a result of uninterrupted rewashing over long periods of time, the material should be carried eventually from higher levels of the bottom, above wave base, into deeper places. Thus the undulatory vibrations hinder deposition on the bottom above a certain level. If during the process of uneven deposition, sediments on the bottom of a basin are piled up higher in one place and lower in another, the undulatory vibrations will level the irregularities, redistribute the material, and transport it from the elevated zones into lower ones. The random vibrations of the water "polish" the bottom, bringing it to a uniform level.

If all the space below the base of wave action is already filled with sediments, then the newly arriving material will be transported, after prolonged washing and rolling, over the boundaries of a shelf on which there is no place left for the stable accumulation of sediments. But as soon as subsidence of the earth's crust causes a deepening of the basin, a certain space will be exposed below wave base; within this space sediments may accumulate and remain. Let us assume that subsidence occurs at an uneven rate in different parts of a basin; the height of the region of possible accumulation would thus be different in different places. In this case, if a sufficient amount of clastic material is available, the accumulation will be equivalent to the subsidence of the earth's crust at every given point.

It is easier to explain by the following reasoning: Let us admit that the earlier existing space for possible accumulation is filled and that afterward for some time delivery of material ceases. Meanwhile the earth's crust has subsided in some places more than in others. Delivery of material is then renewed. At the beginning it is more or less evenly distributed over the bottom of the basin, and the thickness of the sedimentary bedding initially increases with uniform rapidity. But in areas of less subsidence the surface of the new sediments will reach wave base earlier, and in this area the accumulation will be removed by wave action. The material will be carried over these places by the undulatory vibrations and piled up in deeper areas where accumulation continues. Clearly, this process will terminate when the surface of the sediments everywhere reaches wave base. At this time the thickness of sediments will everywhere be equal to the subsidence of the earth's crust and will change from place to place, corresponding exactly to the change in the amount of subsidence of the marine basin.

We have divided the whole process into two stages: *subsidence* and *accumulation*. In actual fact, subsidence and accumulation of sediments occur simultaneously.

It thus appears that this mechanism provides exact compensation of subsidence by accumulation. The following conditions must be met if the mechanism is to operate: (1) the depth of the wave base must remain unchanged; (2) the undulatory vibrations must have enough

time to distribute the delivered material over the bottom before a new portion available for deposition appears, and the sedimentary material everywhere up to the depth of wave base must have sufficient mobility to be displaced by the undulatory vibrations of the water; (3) there must always be enough material to fill the region of possible accumulation.

Failure to fulfill these conditions will cause the top surface of the sediments to develop at different depths. This surface will be either rising (for example, together with the elevation of the wave base or the deposition of denser sediments less apt to be removed) or subsiding (for example, because of the lack of sediments). Then exact compensation of subsidence by accumulation will not occur; the subsidence will sometimes outdistance and sometimes lag behind the accumulation. If these conditions are not fulfilled, one must determine the degree to which they are not met in order to measure how much this affects the mechanism of compensation.

The wave base is, of course, subject to fluctuation. It is determined by the deepest penetration of waves during the most violent storms. But the intensity of the undulations changes with changing climatic conditions, and particularly with the changes in the configuration of land and sea. Where, as a result of crustal movement, a small closed basin or gulf is formed in place of an open sea, the force of the waves is diminished and the wave base rises. The reverse changes lead to a lowering of the wave base. The force of the waves diminishes as one approaches the coast. Hence the surface of the wave base rises toward the shore; but the dip of this surface can be changed by changing geomorphologic conditions.

Therefore wave base cannot be considered constant. It fluctuates within very small limits; these fluctuations occur between the surface of the water and 200 m (the maximum depth of penetration of waves in the open sea). Consequently, the surface of the sediments at the bottom of a basin may rise or drop, but so far as these fluctuations are due to the fluctuation of the wave base, they can have only a limited amplitude.

The level of the surface of the sediments on the bottom of the basin depends on the speed with which the undulatory vibrations wash and distribute the delivered material. If the material is delivered relatively quickly and crustal subsidence is relatively slow, the washing always will proceed simultaneously with the delivery of new sediments. The surface of the sediments will remain higher than the wave base, because under these conditions it will always lag somewhat behind the washing and redistribution of the material. The position of the surface of sedimentation will be determined by the dynamic equilibrium between the rapidity of delivery of material, the rate of its washing, and

the rate of crustal subsidence. If the rate of delivery of material increases relative to the rate of subsidence, the surface of the sediments will rise; in the opposite case, it will drop. The level of equilibrium will be displaced up and down. What are the possible dimensions of this displacement?

To begin with, it should be noted that the closer the particles of water are to the surface, the stronger are their vibrations and the greater is the effect of those vibrations in hindering the elevation of the surface of deposition. Therefore to raise the surface of the sediments from a depth of 100 m to a depth of 50 m one must have a considerably greater acceleration of the delivery of material or a greater retardation of subsidence than to raise the same surface from a depth of 200 m to a depth of 150 m. But it may happen that the delivery of the material is so great that the undulatory vibrations cannot prevent the surface of the sediments from rising, and certain coastal regions of the basin will become filled. The surface of the sediments rises above that of the water. In this case the equilibrium level will be determined by the interaction between the rate of delivery of material from the more elevated sections of land, the subaerial erosion, and the crustal subsidence. Among the continental facies in the geologic section we find almost exclusively the facies of sediments from coastal alluvial plains, which are only slightly elevated above sea level. Other facies of continental sediments also occur in the geologic section, but they are of relatively small significance.

We can conclude from this that stable accumulation of sediments on dry land usually occurs only in very low places. At a slight elevation above sea level, on the average, from the point of view of geologic time, the processes of disintegration are already in operation. The height of the boundary between the zone of prevailing accumulation of sediments and the zone of prevailing disintegration of the land (the level of equilibrium) drops and rises, as in the sea, depending on the rate of delivery of material, the rate of erosion, and the rate of crustal subsidence. But, judging from the continental sediments found in geologic sections, the level of equilibrium probably does not rise over 100 m above sea level. Sediments deposited above this level are very rarely preserved in the section. Therefore the rate of delivery of material affects the level of the surface of the sediments, but fluctuations in this level caused by delivery of sediments are not very large; they are chiefly between the limits of -100 m and $+100$ m above sea level.

Some irregularities in the surface of the sediments result from the varying ability of different rocks to resist erosion. Coral limestone can sustain a considerably greater agitation of the water than porous sandstone, which is easily disintegrated. Therefore the level of equilibrium between the accumulating and eroding actions of the waves will be

reached at a higher level in coral limestone than in sand, and a coral reef will be elevated above the surrounding arenaceous sediments. The fluctuations of the surface of sedimentation caused by this action cannot be large and, in any case, are within the limits shown above.

If the surface level of sediments is not to exceed the said limits, a sufficient amount of sedimentary material must be delivered for the constant filling of space available for accumulation. At times, the delivery of clastic material from the land may become insufficient to fill this space. In this case, the deficiency is usually compensated by material of different origin—organic or chemical sediments. If such compensation did not occur, the observed maintenance of shallow depths could not take place during crustal subsidence.

At first sight, it is not fully clear why there is sufficient clastic, organic, or chemical material available to keep up with subsidence in a marine basin. Let us note that this is an established fact based on observation, which we must accept, even if we do not understand the reasons causing it. It should not be forgotten that geologic sections contain neritic basins, which are the most favorable places for the development of life. The number of organisms, including rock-builders, in such basins is always large, and we can assume that there always is enough organic detritus to fill that part of the space of possible accumulation which is not filled by clastic material. Moreover, organic and chemical material, like the clastic sediment, is subject to rewashing and redistribution at the bottom of the basin, and this process helps to distribute the sediments to correspond to the scale of crustal subsidence.

Hence our explanation of the general correspondence between subsidence and accumulation acquires a better foundation. The reason for this correspondence is considerably more complex than crustal subsidence under the weight of sediments; it includes the rewashing of the sediments by undulatory vibrations of the water, usually within a shelf basin having a sufficient quantity of sedimentary material. The deciding role is played by the wave base. We have also outlined the reasons for the small anomalies diverging from perfect compensation of crustal subsidence by accumulation, as shown in the study of particular sections. These discrepancies are due to changes in the rate of subsidence, the depth of wave base, the rate of delivery of the material, its different susceptibility to wave action, and perhaps, at times, to the lack of sedimentary material. The divergence from exact compensation caused by these factors is relatively small—possibly not over 300 m.

The discussion until now has dealt with uniform crustal subsidence, but its movements are actually very complex. Regions of uplift exist near zones of subsidence, and wherever subsidence occurs, it is generally interrupted from time to time by opposite motion of a lesser order, complicating the subsidence and making it irregular.

Let us consider the following situation. If crustal uplift follows subsidence and accumulation of sediments, then the accumulated sediments will rise above the level of equilibrium and the thickness of the accumulated sediments will decrease during the crustal uplift. With new subsidence the accumulation reappears. If the uplift develops quickly, the top surface of sedimentation may rise above sea level before the sediments are disintegrated by wave action. In this case, the thickness of sediments resulting from the preceding subsidence will be largely preserved. It is the present author's intention, however, to clarify the conditions leading to the development of the thickness of strata that remain fixed in the section, that is, those which are covered by later sediments. In this case, the *later* strata can be deposited only when the surface of sedimentation again subsides below the level of equilibrium, and the increase in their thickness will correspond to the crustal subsidence below this level.

Therefore the final thickness of the deposits, under the conditions of complex oscillatory crustal movement with intermittent subsidence and uplift, will be equal to the algebraic sum of the motions of the earth's crust. Every subsidence causes a corresponding accumulation; every uplift a corresponding erosion, or at least a suspension of accumulation. If subsidence predominates during intermittent subsidence and uplift, the sum of motions is negative and the net crustal movement is subsidence; thus the thickness of sediments will increase in accordance with the resultant subsidence, the algebraic sum of the separate fluctuations. It is absolutely immaterial how many of these fluctuations occurred, how great their amplitude was, and how they have been distributed in time.

Let us assume that in a certain region crustal movements occur with predominant uplift and, as a result, the region is continually and stably elevated. Subaqueous and later subaerial erosion gradually diminish the thickness of the deposits accumulated earlier. In this case, the preserved thickness is of no interest to us, as it is not yet fixed in the section. The new deposit, fixing the thickness in the geologic section, may occur either during full subsidence of the land below the level of equilibrium or during marine erosion, which cuts it off. In the last case, the surface of the earth's crust is again reduced to the level of equilibrium, as determined by the position of wave base and the other factors listed above. During erosion from the continent there remains the leveled top of a pedestal, the surface of which in the ideal case coincides with the base of wave action, and a shelf is formed. All regions situated above the level of equilibrium are cut down. And on this shelf plain, leveled by the undulatory vibrations of the water, the accumulation of sediments reappears during additional crustal subsidence.

In this abrasion and cutting down of the continent to a leveled

surface we may see the reason why transgressive strata are almost always initially deposited horizontally and why geologic sections as a rule do not preserve the relief of the former dry land.

Let us give an example. Assume that in a hypothetical region, during an earlier subsidence, a deposit of sediments 1,000 m thick was accumulated and that the surface of sedimentation is situated near wave base. Then the crust was uplifted 800 m. Before the deposit had time to be measurably destroyed subsidence recurred, accompanied by a transgression of the sea and its erosive effect; the same deposit was eroded down to wave base, and the crust subsided 300 m before the erosive action was completed. Figure 95 shows a diagram of these motions. After a subsidence of 1,000 m, uplift of 800 m, and new subsidence of 300 m, the crust has sunk 500 m from its initial position. Hence this is the figure determining the final thickness of the sediments. If the level of equilibrium does not coincide with the wave base, the final preserved thickness will then be different from 500 m, but the small difference is negligible. This example quite clearly shows the mechanism of the formation of sediments fixed in the geologic section.

An exception to the established rules is the development of thick ancient continental deposits, formed in closed basins, where material carried in from surrounding heights cannot be transported farther away. In this case, the thickness of the accumulated sediments will correspond to the amplitude of local subsidence of the crust only where such deposits are covered by marine sediments. Divergences from the same rule are also observed where an ancient buried relief is preserved. All these are merely exceptions to the usual character of the process.

The given rules are not applicable to oceanic sediments in the great depths beyond the margins of the continents, because the space available for accumulation cannot be filled where insufficient sedimentary material exists. Up to the present such deposits have not been found in geologic sections, and probably they do not exist in the modern continents.

It must be stressed that correspondence between crustal subsidence and accumulation can be established only for a prolonged geologic time. To bring about such correspondence, rewashing of the sediments and leveling of the surface of sedimentation during rewashing are necessary. In any short segment of time rewashing is in considerable degree orderless, and only after a very long time does it produce statistically uniform results.

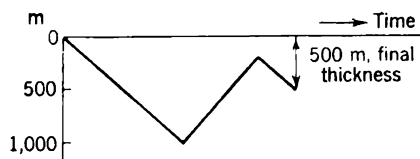


FIG. 95. Diagram of a complex oscillation of the crust.

Thus the difference between the accumulation of recent sediments and ancient deposits fixed in the geologic section becomes understandable. The deposits now forming on the bottom of a basin either are not yet subjected to rewashing and redistribution or else those processes are still continuing. Recent sediments are found chiefly where they were originally delivered, and their quantity depends on the local geomorphologic conditions. But gradually, in the process of constant and complex oscillatory movements of the crust, with the aid of undulatory currents and subaerial erosion, these sediments are subjected to rewashing, which finally causes the thickness of the sediments to correspond to the amount of crustal subsidence. During the process of redistribution, the influence of the local conditions determining the initial thickness is governed only by the general factor applying to all sediments: crustal subsidence.

Therefore the factors controlling the thickness of recent sediments and the factors that control the thickness of deposits in the geologic section are different; more exactly, their relative significance changes with time. The factor of crustal subsidence also acts upon the accumulation of recent sediments. But under recent conditions, where we are dealing with a short segment of geologic time, this slowly acting factor remains imperceptible. By way of contrast, the inconstant, changing, but violently acting physicogeographic factors, which obliterate the direct influence of crustal movements on the thickness, command our attention. This in no way testifies to the peculiar nature of the modern epoch, nor does it prove that in the past geologic processes occurred differently. In any short segment of geologic history, the same picture as now obtained on the bottom of marine basins and the thicknesses of newly formed sediments initially depended on the local physicogeographic conditions. But with the passing of time the sedimentary material was redistributed, its primary disposition was changed, and at the time when crustal subsidence fixed them in the geologic section, the distribution of thicknesses was already subject to the more general factor of crustal subsidence.

Despite the fact that our direct observations of the process of sedimentation in modern basins are extremely sketchy, the existence of the phenomenon of rewashing of the sediments can be directly demonstrated: storms always carry bottom silt and sand into shallow places; after a big storm the configuration of the bottom is sometimes considerably changed, and cases of displacement of the clastic material along the shore, as a result of the obliquely striking waves, are observed.

MECHANISM OF FORMATION OF LITHOLOGIC FACIES

Which processes cause the distribution of the lithologic facies of sedimentary rocks observed in the geologic section? In solving this problem,

we will only be interested in the final distributions of the facies, those that are fixed in the geologic section. It is evident from the preceding paragraphs that the final thickness does not correspond to the initial distribution. We shall establish something analogous for the facies as well. We shall speak, as above, of the most general rules and only note divergences in passing. Naturally, the formation of facies in the sediments will interest us only so far as it is determined by tectonic factors.¹

The earth's surface may be divided into zones of erosion and zones of deposition. The zonal boundaries often coincide with the shore lines, but not everywhere. The same boundaries may be found on dry land where the zones of erosion and accumulation of continental deposits touch each other. They may occur below the surface of the sea, because of underwater erosion. Henceforth we shall speak not of the shore line, but of the *boundary of erosion*, making this concept more precise where necessary.

Sorting of clastic material according to grain size occurs in the zone of accumulation: the coarsest material is deposited close to the boundary of erosion; farther away the finer-grained material is deposited. Deviations from this rule commonly occur in recent basins of deposition; argillaceous sediments deposited on the stagnant bottoms of bays may be replaced by sand bars toward the open sea. These deviations are induced by the complex shoreline, by the distribution of offshore currents, by geomorphologic factors, by the local delivery of material of a specific grain size, by the environmental conditions of rock-building organisms, etc. The influence of these local factors, often changing within short distances, sometimes leads to a very great variety in the facies of recent deposits.

In the sedimentary rocks of the geologic column, variety in distribution of facies, although present, is of subordinate importance. In general, the geologic section is characterized by homogeneity of facies over great areas (see Figs. 90 to 94). Layers of conglomerate, sand, shale, or limestone can be traced, almost without change, for tens and even hundreds of kilometers in all directions. The black shale in the Volga region covers many tens of thousands of square kilometers. Limestones of the Upper Devonian and Carboniferous systems may be traced throughout an even greater area in the eastern part of the European U.S.S.R. The local development of a particular sediment is much rarer. Farther from the ancient zone of erosion the character of the sediments changes regularly; from the maps, cited above, one can see that coarse sediments accumulate near the boundaries of erosion, the grain size gradually but

¹ By *facies* we mean here the composition of the sediment, indicating its position relative to the eroded area. From this point of view the following simple sequence of facies could be established: gravel, sand, clay, and calcareous sediments, indicating, in this order, an increasing distance from the area of erosion.

regularly changing from the center of the zone of accumulation into finer-grained and finally to organogenic sediments.

By analogy with what was said above regarding the development of sediments of varying thickness, it seems that the regular distribution of facies and the broad areal distribution of homogeneous sediments result from some process that has not had time to affect recent deposits. To test this hypothesis, let us examine the mechanism of sorting of sedimentary material in the zone of deposition. The cause of sorting is evident where sediments are deposited on dry land by flowing waters; streams flowing down the slope decrease their velocity, kinetic energy is lost, and their transporting capacity diminishes; consequently, clastic material settles on the bottom in the order of decreasing particle weight. In considering geologic conditions, it is clear that the conditions of stream flow are variable, especially on low alluvial plains. The stream channels migrate from place to place, causing continental sediments to be deposited where the stream ceases to flow. In time such a stream, gradually moving from one side of its flood plain to the other, can cover a large area with its sediments. We must see in this the reason for the very wide, uninterrupted distribution of certain ancient deltaic and alluvial deposits. For instance, the Ufa and Tatar deltaic deposits on the western slopes of the Urals, extending over hundreds of thousands of square kilometers, should in no way be considered as the sediments of a single delta of a certain water artery of unknown dimensions. In fact, many ordinary streams flowed down to the west from the Urals range across the coastal plain, leaving their sediments in a comparatively small area. But the river channels migrated across the plain, gradually distributing these sediments over a large area.

Initially the sedimentary material was deposited unevenly, and the distribution of its facies was irregular; the coarse and fine materials had not yet been completely sorted and, because of the temporary positions of the channels, were piled in heaps and sinuous bands. But as their channels were displaced, the rivers rewashed their own deposits again and again, correcting the local irregularities in the distribution of clastic material. Such rewashing was promoted by oscillatory crustal movements, displacing the boundary of erosion and transforming the same region from a zone of erosion into a zone of distribution, and vice versa. As the result of prolonged rewashing and redeposition, the primary distribution of the sedimentary material was not preserved. The gradual displacement of the material established a smoother and more uniform distribution of the sediments than initially, so that local and temporary irregularities were reduced or completely obliterated. As a result, regular distribution of material was achieved in wide zones containing sediments of decreasing grain sizes, in accordance with the distance from the boundary of erosion. The accumulation of large thicknesses is

possible only during the subsidence of the crust. As a result of this subsidence, redistributed sediments with an averaged and simplified distribution of facies will be fixed in the geologic sections, after being covered by more recent sediments.

In this process, the rate of subsidence of the crust, or, more correctly, the ratio of the rate of subsidence to the rate of delivery of clastic material, is of considerable importance. If subsidence is relatively rapid, the sediments undergo very little rewashing. In this case, the local irregularities in the distribution of the material due to the temporary positions of the channels will be preserved, to a great degree, because the sediments, quickly buried under new deposits, have not been subjected to more thorough rewashing and redistribution. If subsidence proceeds slowly, the averaging of the distribution of sediments and the leveling of local irregularities are usually more thorough because the sediments, before being submerged under subsequent deposits, undergo more prolonged rewashing. There will be a greater variety in the distribution of facies in the first case than in the second.

Let us consider marine sediments. Clastic material on the bottom of the sea, as well as on land, is sorted according to grain size. But whereas this sorting is done on land by flowing water, on the bottom of the sea it is done by the undulatory vibration of the water particles in waves. In its circular orbit a particle of water moves forward (in the direction of the wind) with greater velocity than backward, so that in its vibrations the water moving forward possesses greater kinetic energy than that moving backward. But close to the shore the movement of the waves is always perpendicular or almost perpendicular to the shore line. Hence the more energetic movement of the water toward the shore brings to the coast all the material available to the waves of a given force, and the less energetic reverse movement of the waves carries the finer material away from the shore into the depths of the basin. This process leads to the gradual fractional sorting of the material.

We see that the causes of sorting on the bottom of the sea are the same as on land: the movement of water changing its velocity and kinetic energy. But whereas on land we are dealing with streams, on the bottom of a sea the motion consists of a multitude of separate short displacements of water particles. In this way the mechanism on the bottom of the sea works constantly with time to level the original irregularities in the distribution of material, bringing the coarse fragments to the shore and carrying the finer farther away, ultimately establishing a regular sequence of facies. Therefore the irregular distribution of facies of sediments on the bottom of modern seas, as described above, must be considered a temporary phenomenon, occurring when the process of sorting and redistribution of the material has not yet been completed.

A knowledge of the conditions determining the extent of the final distribution of terrigenous material on the bottom of a basin is exceptionally important for an understanding of the development of a facies. A comparison of paleogeographic maps shows that the transition from one geologic epoch to another involves a change in extent of distribution of clastic material. In some cases, this material is concentrated only in a narrow zone around the region of erosion, and farther away it is replaced by organogenic sediments; in other situations, these deposits cover the entire bottom of a basin, and no room is left for organogenic deposits. Let us try to clarify the reasons for these changes in the extent of distribution of terrigenous material.

We shall assume that the surface of the already deposited sediments corresponds to the level of equilibrium. Thereupon subsidence of the bottom of the basin takes place, and below the level of equilibrium the volume available for sediment accumulation increases at a certain hypothetical uniform rate. Simultaneously clastic material is carried into the basin. It is distributed by the undulatory water vibrations, and finally all the material will be within the volume available for sediment accumulation. The extent of this region is determined by the rate of subsidence in the zone of accumulation in relation to the rate of delivery of material from the zone of erosion. When the rate of subsidence is constant, increased delivery of material leads to an expansion of the zone of accumulation of clastic sediments and a decrease in delivery leads to a reduction of this zone. Similarly, with a retardation of subsidence, under the conditions of uniform delivery of material, the sediments are distributed over a broader area, and if the subsidence is accelerated, they occupy a smaller region in the volume available for accumulation.

Another value can be substituted for the rate of delivery of material. The intensity of erosion of the land depends on the rate of its uplift, on the distribution of rocks and their resistance to erosion, and on other more or less variable geomorphologic factors. Of these factors, only the first plays an important role. The influence of the erosion resistance of rocks in a locality with a complex structure changes rapidly and is averaged for a short segment of geologic time. The other secondary and incidental factors average out even more quickly. Bearing in mind these general rules and neglecting the possible but insignificant errors, we can substitute for the rate of delivery of material in our ratio the rate of uplift of the erosion zone. Then the extent of the zone of distribution of clastic deposits (x) is determined by the ratio of the rate of uplift of the zone of erosion (h) to the rate of subsidence of the zone of deposition* (s):

$$x = \frac{h}{s}$$

During a decrease in the rate of uplift, the amount of clastic material decreases and is deposited in a smaller zone. But the same result is achieved by accelerated subsidence of a bottom, if the rate of uplift of the land area remains unchanged; then the volume available for accumulation grows more quickly and the same amount of clastic material is concentrated in a narrower zone, correspondingly acquiring a greater thickness.

Expansion of the zone of terrigenous sediments is caused not only by acceleration of the rate of uplift of the land and an increase in the quantity of clastic material but also by retardation of the subsidence in the zone of accumulation. The same volume of sediments, having a smaller thickness, is thus spread over a wider area.

This reasoning is extremely simple, but it contains an element that must be considered with great attention. To permit erosion to act over a long time, an equally long-lasting process of uplift is necessary. This is very often forgotten, so that in establishing either a large or a small delivery of detrital material, one commonly speaks of a high or low relief, considering it to be unchanging. We have already shown that, in the absence of constant uplift, the relief is quickly destroyed. The high or low relief in the zones of erosion should always be understood to result from correspondingly fast or slow rising of the earth's crust. The uplift initiates erosion, and the relief of the surface corresponds to the temporary local dynamic equilibrium between two oppositely acting forces. The more rapid the uplift, the higher the hypsometric level on which dynamic equilibrium is established. Therefore one must always visualize the conjugate processes of erosion and accumulation as acting near each other during uplift and subsidence, that is, during crustal movement, but never in immobile elevated and depressed zones. It is interesting, of course, in paleogeographic studies to map the positions of the mountain ranges in preceding geologic periods. This can be done only qualitatively, by reasoning that the coarse clastic deposits were derived from areas of high relief and fine sediments were usually deposited close to the foothills of a low land. The amount of delivered clastic material itself directly testifies only to the rate of uplift, not to the ruggedness of the terrain.

The relief might be relatively low, regardless of fast tectonic movement, if the rock masses are quickly destroyed and carried away. In the same manner, if a marine transgression is uneven, penetrating the dry land faster and farther in one direction than in another, then as a rule this does not prove that a continent with unvarying relief is subsiding as a whole unit under the surface of the sea. It usually means that the continent, flooded by the advancing sea, is tectonically heterogeneous, that it is broken into zones of crustal subsidence and uplift, and that transgression occurs in the places of subsidence. Briefly, the relief, re-

flected in the geologic section in the form of a separation of the zones of erosion and deposition, exists as long as uplift and subsidence occur. These zones must be regarded as being in a state of motion and not as immobile masses.

It was stated above that where the clastic material fails to fill the space available for accumulation under shelf conditions, there is always enough organogenic material to fill this space. Therefore throughout the zone of clastic rocks we usually find organogenic sediments, chiefly various limestones,² replacing the clastic sediments laterally.

If the ratio of the rate of uplift in the zone of erosion to the rate of subsidence in the zone of accumulation changes, the distance to which the clastic material must be carried from the shore changes correspondingly. The boundary between the clastic and organogenic sediments is displaced to one side or another, either toward the shore or away from it. As a result of numerous such displacements, at the

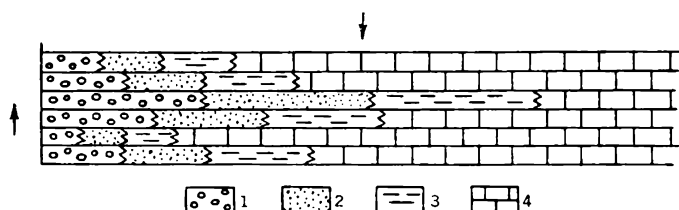


FIG. 96. Interlayering of deposits due to changing widths of facies zones, determined by the ratio of the rated uplift in the zone of erosion to the rate of subsidence in the zone of deposition: (1) conglomerates; (2) sandstones; (3) shale; (4) limestone.

boundary of two facies zones one observes an interbedding of terrigenous rocks with limestones (Fig. 96). But interbedding also occurs in sections of rocks of different composition in the zone of accumulation of strictly terrigenous sediments.

For the sake of simplicity, let us assume that the delivered clastic material consists of three fractions—coarse, medium, and fine—and that the volumes of all the fractions are equal. In the case of ideal sorting, one-third of the width of the zone of clastic material will be occupied by coarse sediments, the second third by medium sediments, and the last by fine sediments. With a change in the zone of accumulation, the width of each part changes correspondingly. What happens during this

² The positions of the clastic and organogenic facies can often be different. All material, including organogenic, is involved in rewashing and sorting. It can happen that the coarse shell fragments will be carried closer to the shore than the zone of living shells. In geologic sections of transgressive and regressive beds, limestone sometimes occupies the place of sediments coarser than, for example, shale. The explanation is similar. Secondary processes of solution, recrystallization, etc., can also change the appearance of the limestone to such a degree that its original coarse detrital texture is not preserved.

process is shown schematically in Fig. 96, from which it is evident that during the changes in the rate of uplift of the zone of erosion and the rate of subsidence in the accumulation zone, we will nearly everywhere observe the interbedding of sediments of different facies.

Thus the observed facies change in the vertical direction in geologic sections has an explanation. The sedimentary bedding in a geologic section is also characterized, above all, by the vertical alternation of sediments of different lithologic composition. Therefore the bedding of deposits can be considered to be the result of constant changes in the ratio between the velocities of crustal uplift and subsidence.³

With the aid of the above formula ($x = h/s$) we can interpret some changes in facies observed in the Russian plain at the transition from one series to another as a result of correspondingly fluctuating tectonic movements. This method provides an explanation for the observed change in facies of the deposits in the geologic section. Let us first note the sharp widening of the zone of terrigenous deposits observed on the Russian plain at the end of the Givetian age, at the beginning of Viséan time and at the beginning of the Middle Carboniferous (the Verey beds). All these can be explained by the temporary retardation of crustal subsidence, without looking for an explanation in a change in the rate of delivery of clastic material. Secondly, by the same temporary retardation of crustal subsidence we can explain the presence of red beds in the Kashir limestone and overlying strata of the Middle Carboniferous in the Moscow basin; as soon as the subsidence of the crust is retarded, the zone of distribution of terrigenous rocks expands, reaching a region where clastic material was not deposited at other times, and the overlap of terrigenous rocks on carbonate rocks occurs over very long distances.

From the same ratio it is possible to explain and interpret the changes in facies observed in the Upper Cretaceous rocks at the northern boundary of the Donets basin. Here in the Cretaceous a transgression of the sea gradually flooded the folded Donbas. It would seem that the sinking of the surface, which caused the marine transgression, should result in a decrease in the rate of delivery of clastic sediments. In reality, however, a ring of clastic sediments bordering the Donbas on the north (see Fig. 94) gradually widened during the Late Cretaceous (Shatskiy, 1924). This process was completed in Paleogene times, and the entire Dnepr-Donets region became an area of deposition of terrigenous sediments.

It may be suggested that these facies changes were caused by gradual retardation of crustal subsidence north of the Donbas during the Late

³ It will be shown below that bedding, and a general change in the nature of the sediments in the vertical section, may also occur under somewhat different circumstances.

Cretaceous. This is indirectly confirmed, first by the fact that at the beginning of the Paleogene there was uplift, so that Paleogene sediments ceased to be deposited, and second by a comparison of the thicknesses of the Upper Cretaceous and the later Paleogene, Eocene, and Oligocene deposits in the Dnepr-Donets region; the considerably smaller thickness of the latter group indicates slower crustal subsidence in the Paleogene. Probably this retardation did not occur precisely between Cretaceous and Tertiary time but developed gradually during the Late Cretaceous.

The distribution of sediments by undulatory water vibrations is the final result of multiple displacements of the sedimentary material along the bottom in different directions. Therefore this process necessarily involves partial mixing of the clastic material of different fractions, with an alternation of sedimentary deposits of greater and less homogeneity and a gradual transition from one facies to another. As a result, the geologic sections, including their facies variations, are more homogeneous than sedimentary deposits laid on the bottom of a recent sea. In the geologic section we notice not only a more regular distribution of the facies but also a wider distribution of homogeneous facies within the area.

Hence the distribution of recent sediments on the bottom of a sea is a temporary phenomenon. Before these sediments can form a stable accumulation fixed in the geologic column, they are subjected to re-washing, redistribution, and displacement. The difference between the distribution of facies in the geologic section and on the open bottom of a recent basin is not to be explained by the supposition that sediments were deposited differently in the past but by the fact that the final distribution of facies is subject to the influence not only of primary deposition but also of subsequent rewashing and redistribution. In the modern deposits we see a unique and temporary picture; in the geologic sections the picture is the net result of the sum of many complex processes acting over a long geologic time. In these processes the phenomena of sorting (splitting into fractions, mixing, and reuniting of the fractions) are combined.

The degree of rewashing and remixing of the material depends on the duration of the rewashing: the slower the crustal subsidence relative to the rate of delivery of the material, the longer the material remains above wave base, so that it is reworked longer and the mixing is more complete. We may expect, therefore, that the zones with a small accumulation of sediments will be distinguished, as a rule, by a greater uniformity in the geologic section than is observed in the zones of sediments with considerable thickness. In fact, the facies on the Russian plain in the Mesozoic, when thicknesses were relatively small, are characterized by greater uniformity than, for instance, the facies

of the Caucasus in the Mesozoic. In the first case, the differences between the facies are usually small, the facies zones are indistinct, and the transition from one facies to another is very gradual. In the second, in the zone of major crustal subsidence and great thicknesses of sediments, the facies differences are usually greater and the boundaries between them sharper.

The same reasoning explains the fact that pure quartz sands predominate on platforms, whereas in geosynclines the sandstones commonly have a mixed or graywacke character. The quartz content of the platform sands testifies to prolonged rewashing, during which the other minerals were in considerable measure disintegrated and dissolved (Rukhin, 1948).

It should also be recalled that subsidence of the crust is never a simple movement in one direction but is always fluctuating. Hence the average rate of crustal subsidence is the algebraic sum of movements

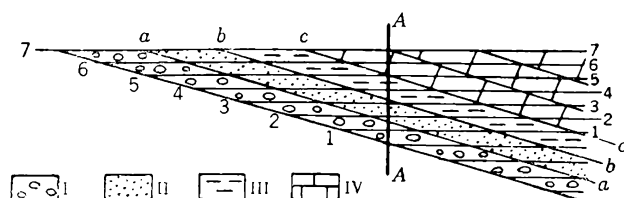


FIG. 97. Displacement of facies zones parallel to the shore line during marine transgression: (I) gravel; (II) sand; (III) mud; (IV) limey mud; *a*, *b*, *c*, boundaries of facies; 1-7, time boundaries; A-A, section (stratigraphic column).

in different directions: the descending oscillations that form the background of the process, and the superimposed ascending oscillations. These ascending oscillations may be of different intensity in different places, causing local alternations in the intensity of rewashing of the material and its distribution, and consequently local differences in the character of the final facies.

The complex factors affecting the deposition and consolidation of marine deposits are not limited to those in this discussion. During crustal oscillations there is a constant redistribution of land and sea, and a constant displacement of the shore lines. If the ratio of the rate of uplift in the zone of erosion to the rate of subsidence in the zone of accumulation meanwhile remains unchanged and only the boundaries of these zones are displaced, then the width of the facies zones remains the same, but the zones themselves are displaced parallel to the migrating shore line. This displacement results in deposition of certain facies above the earlier sediments of differing facies. The vertical alternation of facies may occur any number of times.

Let us examine this process. In Fig. 97 the hypothetical initial posi-

tion of the surface of the sea is represented by 1. The sediments accumulate on the bottom in the usual sequence: gravel close to the shore; sand, mud, and limey mud farther out. Simultaneously with the subsidence of the crust in the zone of accumulation, the shore line moves into position 2. As the width of the individual zones remains the same, all the zones are displaced according to the movement of the shore, and where gravels were once deposited, sand is now laid down, with mud over the former sand and limestone over the former mud. Additional movements of the shore to points 3, 4, etc., continue the process.

As a result, in the vertical section A-A along the shore line, for instance, we observe (from bottom to top) the replacement of gravel

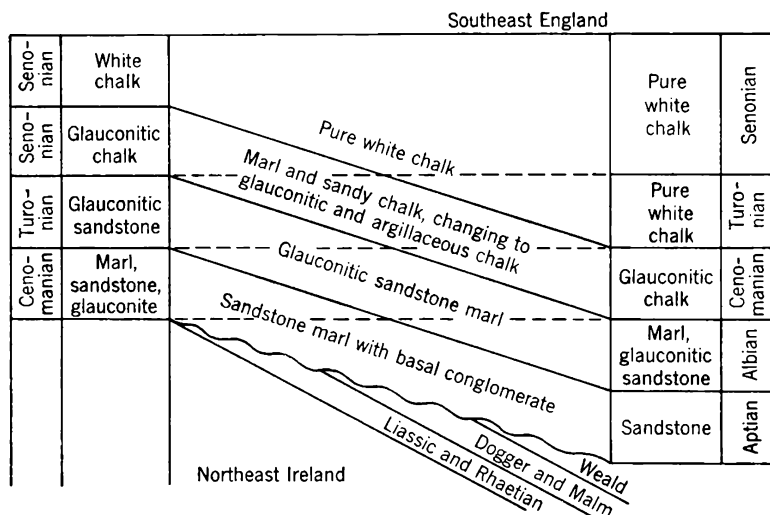


FIG. 98. Migration of facies in England and Ireland (after Grabeau).

by sand, sand by mud, and finally mud by limey mud. This so-called transgressive sequence of sediments proves that with time the shore line moves away from the present location of the section. From the same illustration it is evident that sediments of the same facies in different places are of different ages; in the direction of movement of the shore line the sediments of the same facies are younger. The time-rock units (1-1, 2-2, 3-3, 4-4) do not coincide with the facies boundaries (*a-a*, *b-b*, *c-c*). Thus we observe the phenomenon known as *migration of facies*.

The Mesozoic section in southeastern England and northeastern Ireland in Fig. 98 is given as an example. This shows that in southeastern England the transgression began in the Aptian with the deposition of basal gravel and sand and reached Ireland only in the Ceno-

manian. The deposition of the succeeding sediments in the transgressive sequence was correspondingly later in Ireland: the pure white chalk, crowning the series, was deposited in England during the Turonian and in Ireland only in the Senonian age.

Let us study the case of regression of the shore line. In Fig. 99 the line 1-1 corresponds to the initial position of sea level; on the sea floor sediments are deposited in the recognized sequence: gravel, sand, mud, and limey mud. During the displacement of the shore to points 2, 3, 4, etc., all the facies are correspondingly displaced to the right. In this case, in the vertical section A-A we see a regressive sequence of sediments, starting at the bottom with limey mud and terminating at the top with conglomerate, corresponding to the gradual approach of the shore line to the given point. The difference in the times of deposition of sediments of the same facies in different places is again evident. The large dip to the right of all the formations on the diagram is evidence

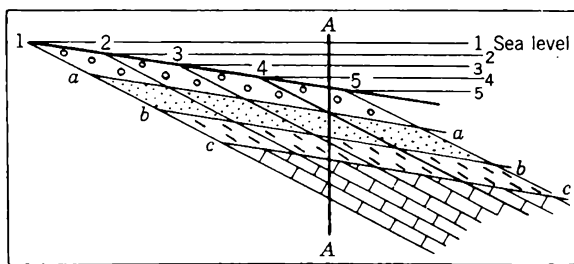


FIG. 99. Displacement of facies zones during regression; legend same as in Fig. 97.

of continuing crustal subsidence in the zone of accumulation simultaneous with displacement of the shore line from left to right. It should be clearly understood that, regardless of the regressive character of the sequence of rocks, their accumulation depends chiefly on crustal subsidence.

Regressive series are not likely to be as well represented in the geologic section as the transgressive, because the final stage of development of a regression is the complete uplift of the crust and destruction of the regressive sediments by erosion. As an example, we may cite the Tertiary section through the Ciscaucasus depression, where the limestones and marls of Paleogene age were succeeded by shale and sand of Miocene age and finally by coarse continental Pliocene sediments. A very typical example is the regressive Permian series on the western slopes of the Urals, where Carboniferous limestones and Lower Permian sands and shales occur below the continental deposits of the Upper Permian.

If the shore line is displaced transgressively and then regressively, a more complex section is created with entire transgressive and regressive

sequences following in vertical succession. By such displacements of the shore in opposite directions and to varying distances, the sequence of sedimentary rocks in the section (Fig. 100) can be explained quite satisfactorily in the majority of cases.⁴

In the preceding examples, we have assumed that the shore line is displaced so slowly that it is easy to detect the dislocation of facies with time by means of index fossils. Examples of this are the changes in the age of the basal strata of the Jurassic transgression from the Ural-Emba region to the Moscow basin, the changes in the ages of the individual facies complexes in the Cretaceous transgressive series in England and Ireland, etc. But the displacement of the shore line may be so rapid that the migration of facies in time is not determinable by stratigraphic methods; it can be established only indirectly. For instance, a widely distributed basal conglomerate occurs in many transgressive series. Fossil occurrences at great distances indicate that these

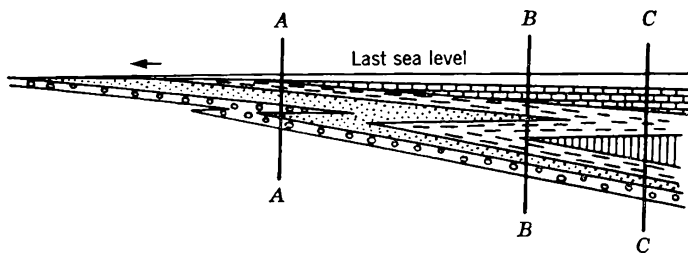


FIG. 100. Migration of facies during complex displacement of the coast line; legend same as in Fig. 97.

conglomerate layers are the same age throughout. An example of this is the conglomerate of Callovian age in the base of the Upper Jurassic transgressive series in the northern Caucasus, extending along the entire northern slopes of the Caucasus range.

In these examples the lack of correspondence between the area of distribution of the basal conglomerate and the width of the strip close to the shore in which pebbles are deposited can be explained only by the fact that the layer of basal conglomerate covering such a large area was not formed at once but rather progressively during displacement of the shore line. The coast line was displaced relatively quickly, and the "carpet" of gravel was extended to keep up with the migration of the shore line. The layer of gravel thus represents a record of the displacement of the shore line. The conglomerate was deposited in the narrow coastal zone, but the zone of accumulation of gravel migrated with the migration of the shore line and earlier-formed gravel portions of gravel were consolidated and preserved simultaneously.

⁴One must, however, always keep in mind the other features of the physico-geographic environment.

In principle, we have the same migration of facies that was determined paleontologically for the first cases. For the Callovian gravel, however, relative age cannot be established paleontologically because the shore line moved too quickly to permit organisms to undergo recognizable change.

It is clear that not only conglomerate but any other widely distributed facies complex of the same age can be regarded as a record of the displacement of the shore line, because all facies zones are displaced jointly with the displacement of the shore. This principle explains the extensive distribution of uniform deposits whose age is not changed noticeably across a wide area.

Our assumption that during the displacement of the shorelines the ratio of the rate of uplift in the zone of erosion to the rate of subsidence in the zone of accumulation does not change is not necessarily true, and it is even arbitrary. In reality, the development of the geologic section is much more complicated, involving simultaneous displacements of the shore line and changes in this ratio.

The following situations may arise:

1. During an advance of the shore, the relationship between uplift and subsidence changes, with subsidence predominating. The facies zones are displaced with the displacement of the shore line, and the zones of terrigenous facies become narrower. Figure 101A illustrates a section through a transgressive series of sediments.

2. During an advance of the shore, the ratio of uplift to subsidence changes in favor of uplift. The zones of terrigenous sediments widen, and therefore at least some of the outer boundaries between the facies zones can be displaced in the opposite direction relative to the shore-line movement. In this case it is quite possible to find an apparently regressive series of sediments preserved in the section, although the sea has advanced toward the shore and the shore has withdrawn from the place of observation (Fig. 101B). The northern margin of the Donbas in the Upper Cretaceous is an example.

3. During a retreat of the shore, the ratio of uplift to subsidence changes in favor of uplift. The facies zones are displaced in the direction of the shore line's movement, and the zones of terrigenous sediments widen. A regressive series of deposits is preserved in the section (Fig. 101C).

4. During a retreat of the shore, the ratio of uplift to subsidence changes in favor of subsidence. The zones of terrigenous sediments become narrower, and some of them may be displaced in the direction opposite to the movement of the shore (Fig. 101D). One may thus observe that transgressive sediments are preserved in the section, although the shore has retreated.

The presence in geologic sections of situations similar to those described in paragraphs 2 and 4 compels us to consider carefully the in-

terpretation of transgressive and regressive series in determining the direction of movement of the shore line; a "regressive" series may originate during an advance of the sea and "transgressive" beds during its retreat. Therefore the terms "transgressive" and "regressive" should be used only to indicate a relatively transgressive or regressive sequence

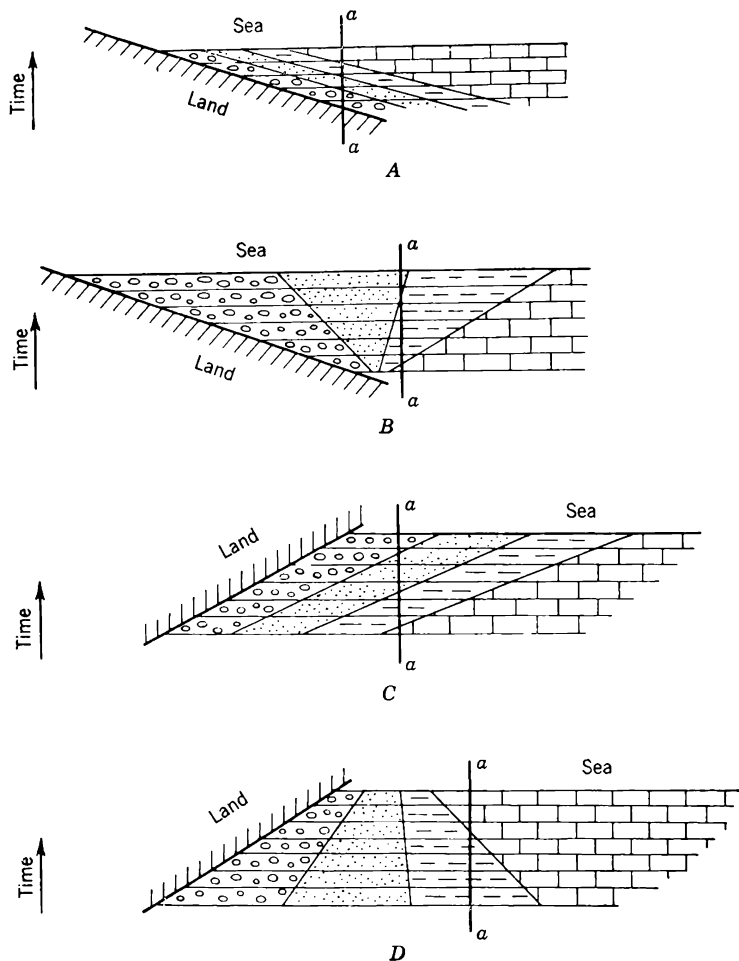


FIG. 101. Widening and narrowing of facies zones during advance and retreat of the sea.

of facies. In terms of the movement of the shore line (the boundary of erosion), it is more appropriate to speak of an expansion of the basin and widening of the zone of accumulation during an advance of the shore and of a reduction of the basin and narrowing of the zone of accumulation resulting from retreat of the shore. The situations mentioned

in paragraphs 1 and 3 are transgressive widening and regressive reduction. In paragraph 2 we have a regressive widening, and in paragraph 4 a transgressive reduction.

An additional illustration is provided by the Upper Jurassic beds of the Kislovodsk region in the northern Caucasus (Fig. 102). South of Kislovodsk the Oxfordian and Kimmeridgian limestones grade into lagoonal deposits of Tithonian age (red beds, gypsum, dolomite). In the vicinity of Kislovodsk the pre-Tithonian strata are absent, and the lagoonal and continental deposits lie directly on ancient granites; that is, during the Late Jurassic the zone of accumulation widened to the north with a regressive change in facies.

Transgressive reduction occurred on the Russian plain at the contact between beds of Early and Late Cretaceous age; the chalky marls and limestones of the Upper Cretaceous are the facies of relatively deeper-water sediments than the terrigenous sediments of the Lower Cretaceous, but the area of distribution of Upper Cretaceous rocks is, and

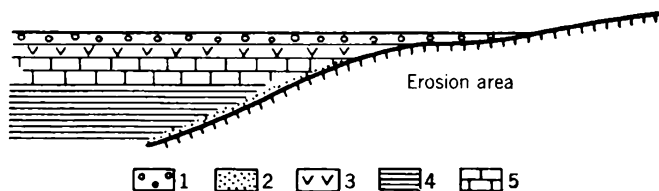


FIG. 102. Diagram of distribution of Jurassic deposits in the Kislovodsk region: (1) red sandstone, (2) marine sandstone, (3) gypsum, (4) clay, (5) limestone.

the area of the zone of their accumulation was, considerably smaller than the area on which the Lower Cretaceous sediments accumulated. Therefore, regardless of the widening of the zone of erosion and reduction of the zone of accumulation, the quantity of terrigenous material and extent of its distribution are sharply decreased; that is, regardless of the reduction of the zone of subsidence, the ratio of the rate of subsidence to that of uplift has changed in favor of subsidence.

Expansion of the zone of accumulation in one direction is not necessarily accompanied by its expansion in others. Displacement of all the zones of accumulation in one direction may occur simultaneously with their expansion in one direction and reduction in another. Similarly, transgression in one part of a basin can take place at the same time as regression in another. For instance, during the Early Carboniferous, the Urals and central Kazakhstan were part of a single basin of deposition. But although in Kazakhstan regression and contraction of the sea were in progress at this time, resulting in exposure and drying of the sea bottom at the end of the Early Carboniferous, in the Urals the coal-bearing sediments of Tournaisian and Lower Viséan age grade upward

into limestones of the Upper Viséan; that is, transgression was taking place.

During the Late Cretaceous and the Paleogene in the Russian plain, at the northern margin of the marine basin, a reduction in size of the sea proceeded inexorably; the sea retreated and the land widened. Simultaneously, farther south, at the northern boundary of the Azov-Podolian massif, the basin expanded, advancing toward and eventually inundating the dry land from north to south. Hence the whole basin was gradually displaced southward. Reduction in the north and expansion in the south were each accompanied initially by a transgressive change in facies (from terrigenous sediments of Early Cretaceous age to Upper Cretaceous limestones and then by regression (from the Upper Cretaceous limestones to terrigenous Tertiary sediments). This terminological usage, justified by the needs of our literature, is indeed unusual. Therefore in the following parts of this text, where possible, we shall use the conventional terms *transgression* and *regression*, defining them as advance and retreat of the sea (that is, as expansion and contraction of the zones of sedimentation). But where it is necessary to distinguish expansion and contraction of the zones of accumulation from facies changes, we will use the new terminology and note the usage. We have stated that the development of facies in sedimentary deposits is analyzed here only from a tectonic standpoint. All the geologic material persuades us that the tectonic factor plays the leading part in the distribution of a facies zone. In any case, it determines the basic pattern of such distribution, marking the disposition of terrigenous and nonterrigenous facies and of clastic rocks composed of sediments of different grain sizes.

Nevertheless, it is unfortunately true that factors of nontectonic character also influence the distribution of facies. For example, changes in the temperature and salinity of the water inevitably affect the sediments, independently of tectonic conditions. These factors to a considerable degree determine the horizontal and vertical changes in limestones and various other organogenic rocks. They also influence the distribution of chemical sediments, although it should be borne in mind that an increase in the salinity of a marine basin, leading to large-scale precipitation of salts from solution, may be determined by tectonic conditions, such as the transformation of an open basin into a closed one as a result of crustal movement.

Windborne volcanic ash causes the appearance of volcanic sedimentary facies (tuffs and tuffites).

Seasonal climatic changes, influencing the scale of transportation of clastic material from the land, are reflected in changes in the character of sediments, resulting in many cases in a development of extremely thin seasonal laminae, or varves.

It must also be noted that the character of clastic deposits depends considerably on the composition of the source rocks. If the transported clastic matter came from a source area consisting chiefly of shales, then naturally we cannot expect the deposition of conglomerates and sandstones. Even directly adjacent to the shore, only argillaceous material will be deposited. The erosion of acidic igneous rocks commonly results in the deposition of siliceous sediments and the erosion of limestones in precipitation of chemically redistributed calcareous sediments.

Hence phenomena of different orders are involved: the tectonic factor determines the most general aspects of the development and distribution of lithologic facies. The details of sedimentation are controlled to a considerable degree by nontectonic factors. These factors, as applied to actual geologic sections, are still very slightly known, although the peculiar features of the process of facies development produced by them are of great economic importance.

THE FORMATION OF STRATA

Stratification is a characteristic of all sedimentary rocks; there should therefore be some general mechanism controlling the development of sedimentary beds.

In analyzing the process of stratification one must distinguish two phenomena: the change in lithologic composition of rocks in a vertical sequence, and the formation of surfaces of separation between the layers. The latter is no less important than the first: very often we separate a series of almost completely homogeneous lithologic composition into different beds. But even in heterogeneous layers the bedding surfaces have a significant meaning; they are determined by the fact that the lithologic composition usually changes discontinuously from layer to layer.

The development of stratification as the result of vertical change in the lithologic composition of the rocks is subject to phenomena that were analyzed above, in the discussions of the change in facies. We have said that the change from one type of sediment into another type of different facies arises from the change in the ratio between the rate of subsidence in the zone of accumulation to uplift in the zone of erosion, and also from the displacement of the boundary of erosion (the shore line). The fact that in bedded rocks there is very commonly a rapid alternation of beds of different composition indicates a rapid change in tectonic conditions; that is, the relationship between uplift and subsidence changed rapidly, as did the location of the shore lines.

Both factors causing stratification act jointly, but one of them predominates. When we see a layer of conglomerate extending tens or hundreds of kilometers, we know that displacement of a shore line has

taken place. In similar cases, what was said previously about differences in time and place in the formation of rocks of the same facies complex and the lack of coincidence of lithologic and time boundaries is fully applicable to bedding. Each bed is such a facies complex. If the bed is the record of the displacement of a shore line, it has been formed gradually as the shore line moved. But because the layer represents relatively short-lived conditions, the time elapsed during the deposition of thin layers in the majority of cases cannot be distinguished paleontologically. An example of development of sedimentary bedding in recent times connected with movement of the shore has been observed on the coast of Panama (Fig. 103). It was found in drilling that the sand and peat forming the shore at the surface could be traced as inclined layers under the sea. Evidently these layers were formed gradually, during an advance of the sea. Throughout the time of

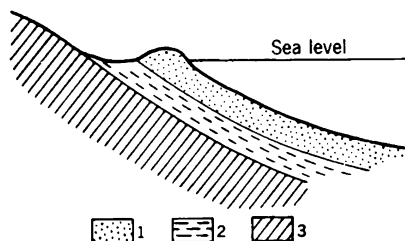


FIG. 103. Formation of modern beds along the shore: (1) sand, (2) peat, (3) basement rocks.

deposition of these layers, the topography of the coastal zone remained essentially as it is at the present time; the swampy area where peat was formed was separated from the sea by a sand bar (McDonald, 1920).

If the stratification is determined mainly by the change in the relative rates of subsidence and uplift, the problem of the correspondence or lack of correspondence in time of deposition of different parts of a

sedimentary stratum becomes more complex. The formation of sedimentary layers takes place by a process of erosion and redeposition of the material, which can be terminated in different places at different times. But a bed can be considered as having been formed simultaneously, in terms of the margin of error allowable for determination of its geologic age.

N. B. Vassoyevich (1951), who studied the problem of stratification in detail, also came to the conclusion that two types of bedding exist, of which one is called *mutational* and the other *migrational*. By the latter he meant bedding formed during a displacement of the shore line. The mutational type of stratification, according to N. B. Vassoyevich, is determined by changes in the tectonic or other conditions occurring while the shore remains in a stationary position. The same author noted that these two types of bedding are seldom encountered in their pure form but are usually manifested in combination.

Yu. A. Zhemcluzhnikov (1950) distinguished between beds and strata. The bed is a formation deposited at a specific time, whereas the stratum is formed gradually during displacement of the shore.

As regards the surfaces of separation between beds, Barrell saw in them the result of a nonuniform subsidence of the crust, which is sometimes accelerated and sometimes ceases or reverses its direction (Barrell, 1917). Whenever retardation of subsidence occurs, the level of equilibrium (the wave base) remains in the same position for some time in relation to the sediments accumulating on the bottom of a basin. Above this immobile surface, washing and redistribution of the material take place, gradually smoothing the surface of the sediments lying below this level. There is time for this surface to be subjected to subaqueous erosion, and the accumulated sediments are compacted and undergo other diagenetic changes. As a result, when the subsidence recurs, the surface, which for a long time coincided with the level of equilibrium, becomes fixed in the section as a surface of separation between the beds. Barrell suggested that the intervals of time in which such interruptions in accumulation occurred may be considerably larger in sum than the time of deposition itself.

Probably the tectonic factor is the major one in the development of bedding planes between the layers. But N. B. Vassoyevich (1951) properly indicated that the interruption can be caused by other rapid changes in the conditions of sedimentation, such as changes in the climate or the properties of the water. He also pointed out that, even with a continuous change in the dynamics of the medium (for example, retardation of the movement of water), the conditions of deposition of sediments change discontinuously, and this, in turn, can produce a well-defined separation between layers.

CONCLUSIONS

In summarizing everything that has been said previously about the deposition of sediments, we reach the conclusion that sedimentation is a very complex process in which the primary deposition of sediments is accompanied by rewashing, redeposition, and displacement of the material. We should clearly distinguish two phenomena: deposition, an initial phenomenon of short duration, and accumulation, a prolonged phenomenon fixing the sedimentary material in the geologic section.

Thus the final distribution of the thicknesses and facies, and also the stratification of the deposits in general, reflect the end result of all these movements and shore-line displacements. Therefore a picture, not of some momentary disposition of facies and thicknesses corresponding to a certain instant, but an average generalized picture of an intricate complex of crustal movements, is reflected in the geologic section.

We have seen that the character and distribution of facies in the geologic section depend basically on the position and displacement of the shore lines (boundaries of erosion) and on the ratio of the rate

of subsidence in the zone of accumulation to the rate of uplift of the zone of erosion. The distribution of thicknesses depends on the disposition of the zones of greater and less subsidence of the crust.

In almost all the examples we have been concerned with marine sedimentation. It is not difficult to show that everything said about marine sediments relates also to continental deposits, or at least to those preserved in the geologic section: that is, to the deposits of the coastal alluvial plains. In the sea the level of equilibrium is the chief factor controlling the distribution and redistribution of sediments. This level is determined by the relationship between the wave base, the delivery of material, and the rate of subsidence of the crust. On the land the same level of equilibrium, determined by the erosional profile of equilibrium, the delivery of material, and the oscillatory movements of the crust, also exists. The subaqueous level of equilibrium grades uninterruptedly and smoothly into the level of equilibrium on land, and the erosional action of the water's undulatory vibrations is replaced by subaerial erosion. Therefore we must consider the zone of deposition of continental sediments as a facies zone equivalent to the zones of marine facies, but occupying its own place in the complex of facies zones. Its position is closest to the zone of erosion. In continental deposits, as in marine sediments, because of the change in the ratio of uplift to subsidence, the interbedding of rocks of different composition occurs, and bedding develops. The same changes, as well as the displacement of the boundary of erosion, cause a more or less extensive distribution of continental deposits in their proper area or their alternation with marine sediments at the boundary of the facies zones.

The thickness of the deposits (if small errors are neglected) is determined basically by the net subsidence of the zone of accumulation. The lithologic facies, however, are determined by much more complicated factors, including the reciprocal action of both local and external processes. The combination of different factors can lead, in many cases, to similar results (for example, the displacement of a facies zone during displacement of a shore line or a change in the ratio of subsidence to uplift).

We have discussed the general consistencies in the deposition of sedimentary layers and some of their deviations, and we shall not return to this problem. We shall only stress that the presence of deviations in no way diminishes the importance of the overall consistency. Any natural process is the result of the reciprocal or mutual action of many processes, each of which may have a particular direction not corresponding to the resultant direction of all the processes. Our approach to the general and to the particular in each case should be determined by whether we are studying a phenomenon as a whole, or only its details.

CHAPTER 14

METHODS OF STUDYING OSCILLATORY MOVEMENTS

The methods of studying contemporary and recent oscillatory movements have been described in Chap. 11.

The previous chapter pointed out that oscillatory movements of the earth's crust that took place in past geologic periods are reflected in the distribution of facies and in the thicknesses of sedimentary rocks; hence the history of these movements must be based on a study of the thicknesses and facies of the sediments. It should be added that oscillatory movements cause stratigraphic breaks, erosion of previously deposited sediments, and displacements of shore lines and boundaries between zones of erosion and deposition. The study of these latter phenomena is thus also included among the methods of investigating oscillatory movements of former epochs. A complete picture of the oscillatory movements is obtained only by a combined study of facies, stratigraphic breaks, transgressions and regressions, and thicknesses.

THICKNESSES

The thickness of accumulated and lithified deposits in the stratigraphic section corresponds roughly to the amount of subsidence of the earth's crust. For greater accuracy, one must also take account of the changes in facies, which indicate the changes in the surface of deposition during the period of time under consideration. If the depth at which the sediments were deposited was the same at the initial as at the final moment of deposition, so that sedimentary accumulation proceeded at exactly the same rate as subsidence, the thickness of the deposits will be an exact reflection of the amplitude of the subsidence of the earth's crust.¹ If the level of sedimentary accumulation rose during this interval of time (as in the formation of shoals), the change in the level of the surface of deposition must be taken into account in determining the

¹ Without allowing for possible secondary changes in the thickness resulting from compaction or squeezing of the sediments, which will be discussed later.

amount of tectonic subsidence from the magnitude of the thickness. If the surface of deposition was lowered (the basin became deeper), the amount by which the surface of deposition was lowered must be added to the thickness of the sedimentary formation. Such corrections for the changes in facies are usually of importance only when the thicknesses being studied are small. It was seen that the change in the level at which the sediments are deposited, to judge by sections through the continents at the present day, does not exceed 200 to 300 m in the great majority of cases. If the formation of shoals or the deepening of the sedimentary basin approaches these figures, considering the normal error in the determination of thicknesses, such correction may for practical purposes be disregarded in dealing with thicknesses of 1,500 m or more. Changes in the level of sedimentary deposition are usually much smaller; for all practical purposes, therefore, in calculations made to the first approximation, no correction need be made as long as the thicknesses are not small.

The exceptions are those cases in which a great discrepancy is observed between the amount of subsidence and the thickness of the accumulated deposits. Such a discrepancy has sometimes been found in geosynclines, or where ridges or monadnocks were preserved on the surface upon which the sediments were deposited, especially if the sediments are continental in character.

Such a discrepancy forces one to use great care in analyzing thicknesses and to try to find similar cases for comparison. These deviations from normal compensatory deposition are both local and temporary phenomena, and they may thus be noticeable only in detailed reconstructions, when thicknesses are studied closely for a small area and a small interval of geologic time.

Deviations from the rate of deposition just sufficient to compensate for subsidence are most often found by comparing the distribution of thicknesses in successive small stratigraphic subdivisions. Since the position of subsiding parts of the earth's crust tends to be stable for lengthy periods of time, special attention must be given to each case in which thicknesses greater or less than those of the rest of the stratigraphic unit differ sharply from the preceding or succeeding unit. But by studying larger areas with less detail and greater stratigraphic time intervals, one can be almost certain of avoiding gross errors, since the irregularities in the distribution of thicknesses will be much less.

The method of studying thicknesses should therefore be considered as one for a general analysis of the history of oscillatory movements, producing an average picture of their development over a very large area and a long time interval. This does not mean, however, that the method cannot be applied to a study of temporary local peculiarities of oscillatory movements. The only difference is that, although this method

may be used "automatically," so to speak, and without great risk to obtain an overall picture of the history of oscillatory movements over a large area and a long time, in more detailed investigations using this method one must be careful to find all the cases of deviation from a compensatory rate of sedimentation and introduce appropriate corrections.

The question will naturally arise: what size area and what length of time should be studied to obtain the most accurate results when "automatically" using the method of analysis of thicknesses? A definite answer to this question cannot, of course, be given, but the analysis of thicknesses should in every case encompass not one structural form (such as a dome on a platform), but a group of structural forms. As regards the stratigraphic interval, a formation is a suitable unit for areas with small thicknesses and a stage for areas with great thicknesses. One should not go to the opposite extreme, however, of taking too large a stratigraphic interval for calculating the thicknesses, since in this case one might overlook very important changes in the nature of the movements that have taken place during that time.

Any method based on a measurement of thicknesses will be valid only if it uses actually observed magnitudes. All assumptions regarding the amount of sediments accumulated initially and the amount that might have been eroded away are extremely undesirable, since there is no way in which they can be verified.

One should proceed from the thickness of the given stratigraphic unit contained in the section beneath the next overlying subdivision. This observable thickness is determined by the sum of all the oscillatory movements that have taken place from the beginning of the formation of the given unit to the beginning of the formation of the succeeding unit. For example, one might study the thickness of Lower Cretaceous deposits covered by Upper Cretaceous rocks. The thickness of the Lower Cretaceous strata is determined by the summation of all movements occurring in that locality between the beginning of the Lower and the beginning of the Upper Cretaceous. One must deal with the total movements, since movements of the earth's crust are never simple and rectilinear and there are always upward and downward oscillations superimposed on a predominant overall subsidence.

It will now be assumed that the Lower Cretaceous deposits are overlain directly by Neogene sediments. In this case, the thickness of the Lower Cretaceous contained beneath the Neogene deposits is the result of the totality of oscillatory movements that have taken place here from the beginning of the Lower Cretaceous to the beginning of the Neogene. The distribution of these movements in time, however, is unknown. It may be that there was subsidence only during the Lower Cretaceous, after which this part of the earth's crust remained in a neutral position.

Or perhaps the subsidence continued throughout the entire time from the beginning of the Lower Cretaceous almost to the end of the Paleogene along with accumulation of great thicknesses, but just before the beginning of the Neogene there was a great uplift, causing erosion of all the deposits except for part of the Lower Cretaceous. Other successions of events are also possible.

When a lengthy interruption is observed in the deposition of sediments, there are various possible interpretations of the absence of any particular suites in the section: they may never have been deposited at all, or else they may first have been laid down and then removed by erosion.

Where the deposits under consideration are directly covered by Quaternary sediments the present thickness means nothing, because the Quaternary deposits still have no fixed position in the section. The present thickness of original deposits under the Quaternary cover can be regarded only as the minimal possible thickness of these deposits. There is no exception to this, unless there are cases in which the Quaternary deposits already have a fixed position in the section; this could happen only if there had been a tectonic subsidence of a part of the earth's crust that was compensated by the accumulation of the Quaternary sediments. In the Kuban and Tera basins, for example, the Tertiary deposits merge into the Quaternary without a break, thus forming a single series of deposits accumulated under conditions of subsidence of the earth's crust. In this case, the thickness of the Tertiary deposits under the Quaternary may be used for the purposes described in this chapter.

In applying the thickness study method one must also include distortions in the thickness associated with folding. The mean thickness must be used when the beds are greatly thickened at the crests and troughs of folds and thinned out on the flanks.

Certain doubts about the applicability of this method may arise when there are extrusive rocks among the sediments. If the extrusive rock is contained within the sedimentary rock, being both underlain and covered by it, its thickness may be included in the total thickness of the suite being studied. If the blanket of extrusive rock is covered by sedimentary deposits formed at the same depth as those below it, the thickness of the extrusive rock must be omitted from the total. When the sediments above the extrusive have different facies from those of the underlying deposits, appropriate correction must be made for the change in facies. But when the extrusive rock lies on the surface and is not covered by sediments, its thickness may not be used.

Study of the thicknesses of deposits can be used only for areas of subsidence and deposition; the method can give only the amount of subsidence but not that of uplift.

The results of studying the thicknesses may be formulated in various

ways. The figures may be used in such a way as to determine the history of the oscillatory movements at a given point during a certain length of time. One might also obtain a picture of the distribution of such movements over an area during an interval of time. The first instance can be represented on a graph such as those shown previously (see Figs. 85, 86, and 89). The geologic time is indicated on the abscissa and the total thickness accumulated from any moment to any later moment (lower curve) on the ordinate, with the scale beginning at the bottom. Thus one can see the course of the subsidence and, from the changes in the slope of the curve, judge the changes in the rate of this movement. Corrections for change in facies are also made; on the same graph a curve of the changes in the depth of the basin is plotted, and the thicknesses are plotted from this curve.

A graph may be plotted of the rate of subsidence and its change with time. For this purpose each stratigraphic unit (with and without correction for the change in facies) is divided by its absolute duration in time. The average amount of subsidence per unit of time (such as one million years) thus obtained is plotted on a graph, with the rate of subsidence on the ordinate and the time on the abscissa. As an example, Fig. 104 shows a graph of the rate of subsidence for the Mesozoic and Cenozoic eras in the northern Caucasus. This graph should be compared with Fig. 86.

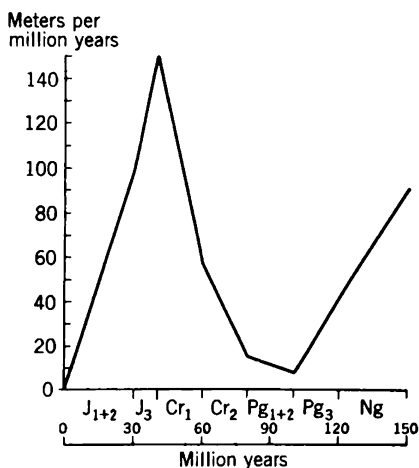


FIG. 104. Graph of the changes in the rate of subsidence of the earth's crust in the northern Caucasus, in meters per million years (after A. B. Ronov).

To obtain a picture of the distribution of the movements during a certain length of time over an area, a map of the lines of equal subsidence of the earth's crust is drawn. If there is no correction for changes in facies, this problem is solved by drawing an isopach map.

The largest possible number of figures showing the thicknesses of the sediments accumulated at various places during the time in question is plotted geographically; the figures used are those for the thicknesses within the section and below the next succeeding stratigraphic unit, particularly that which directly follows the suite under consideration. If, for example, one is studying the oscillatory movements in Lower Jurassic time, one is concerned with the thickness of the Lower Jurassic deposits below the Middle Jurassic in the section. In studying the

movements of the Carboniferous period as a whole, one would examine the thicknesses of the Carboniferous deposits directly beneath the Permian, etc.

Where the sediments under consideration are covered by deposits that do not follow directly in the stratigraphic column, the observed thickness of the sediments must be considered as only the minimal possible amounts and are plotted on the map with the appropriate sign (for example, <80).

Isopachs are drawn, as usual, from the plotted figures. The stratigraphic sections must then be chosen in such a way as to reveal all the main features of the distribution of thicknesses and yet not produce an overly crowded chart. It is generally convenient to plot certain main isopachs on a large section and to insert intermediate isopachs only where the main lines are far apart.

The isopachs must, of course, be plotted by interpolation. If the scale of the map permits it, the isopachs might be smoothed somewhat but this must be done with great care.

The zero isopach will designate the area where the deposits under consideration are absent from the section. The main points for plotting the zero line are, again, those at which the stratigraphic break corresponds exactly to the time interval being studied and at which the section contains deposits formed both immediately before and immediately after this time.

The isopach diagram will give a picture of a region divided into areas where the summation of oscillatory movements has resulted in a net subsidence of the earth's crust (areas with various thicknesses of deposits) and other areas where there has been no overall subsidence (areas of zero thickness). In addition, areas of subsidence will be further subdivided according to degree of subsidence. To obtain a picture of the history of oscillatory movements over a great length of time, a series of successive isopach diagrams must be constructed (see Figs. 90 to 94). These should have sections showing the distribution of thicknesses along several lines. On the section through the top of the suite under consideration will be plotted a horizontal line (if no corrections are made for facies changes) or a curve indicating the distribution of depths in the basin of sedimentation up to the final moment of deposition of the given stratigraphic unit; the thicknesses are read downward from this line. Ordinarily a series of such sections is constructed for the whole succession of stratigraphic subdivisions: a section is drawn for each unit as described above, except that the thicknesses of the preceding stratigraphic subdivisions will be added below those of the unit in question (Fig. 105). In some cases, where the horizontal distribution of oscillatory movements is of little significance or where the

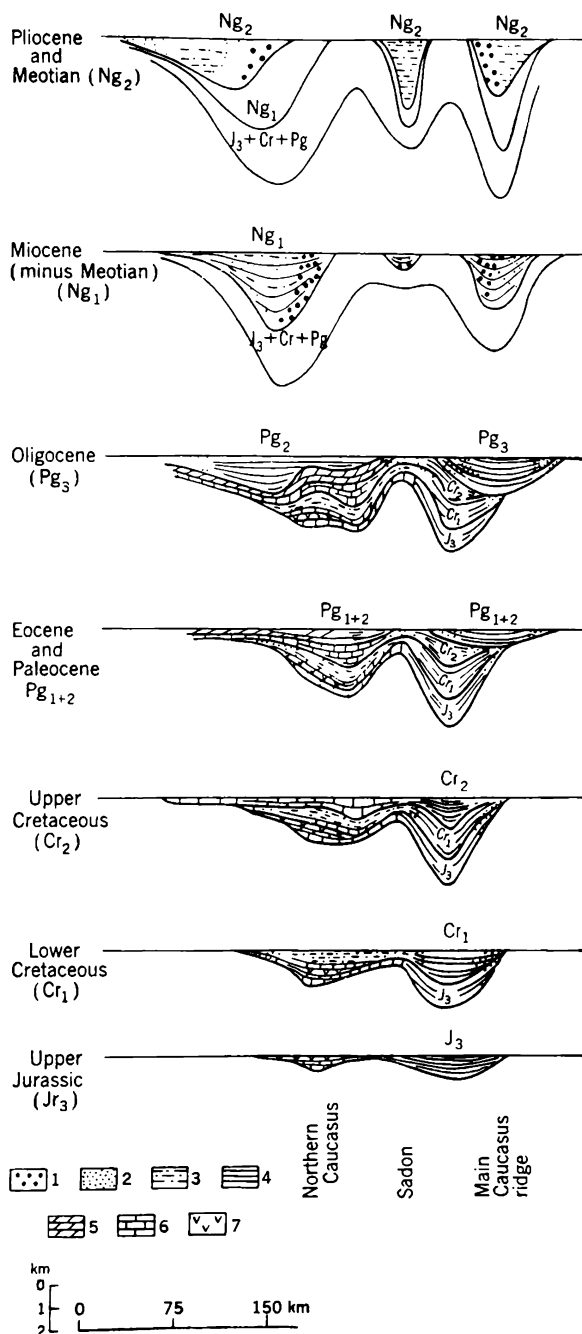


FIG. 105. Series of stratigraphic sections through the Greater Caucasus (eastern part), showing the changes in thicknesses and facies from Upper Jurassic to Pliocene times: (1) coarse clastic sediments; (2) sand; (3) sandy clay sediments; (4) clays and flysch; (5) marls; (6) limestones; (7) lagoonal deposits.

data are lacking, a profile is constructed without drawing an isopach map.

The interpretation of the data obtained in this manner will be explained below; first, however, a discussion of the methods and results of facies studies is in order.

FACIES

By studying the changes in facies from the bottom upward in the section, so as to determine the depth of formation of sediments, it is possible to construct a curve of the oscillations of the earth's crust, that is, of the geomorphologic reflection of these oscillatory movements (see Fig. 84). If this curve is drawn to the scale of absolute geologic time and combined with a curve of the geotectonic subsidence of the earth's crust, the result will be the complete graph of the oscillatory movements figured earlier.

Such an analysis of the stratigraphic section must include a study of all traces of interruption in the formation of sediments or of erosion (stratigraphic nonconformities). On the graph these will indicate the times when the surface of the earth was uplifted above the level of equilibrium.

No other conclusions in regard to tectonic events should be made from such a study of the section. On the basis of the changes in facies in one place, for example, one should not draw conclusions about the approach or retreat of the shore line to or from this locality, because, as has been shown above, the width of a given facies zone can change with time and a transgressive succession of facies can be combined with a decrease in the size of the basin or a regressive succession with an increase in its extent.

It is often expedient to indicate the facies on the profiles of thicknesses mentioned above (see Fig. 105), and thus the interrelationship between the facies and the thicknesses will become clear. A complete analysis of facies is possible, however, only with facies maps showing the distribution of facies zones (see Figs. 90 to 94). To draw these maps, one must determine the different areas in which the facies of a given age were developed; in the general case, one can at least distinguish biogenic, chemical, and clastic facies of various grain sizes. Sometimes the facies of an area change so abruptly that it is easy to pick out distinct zones of different facies; in other cases one facies grades into another so gradually that intermediate zones must be drawn between them or the boundaries must be drawn based on arbitrary criteria. For instance, one might distinguish a zone of sandy argillaceous rock with a predominance of clay grading into a zone of similar rock with a predominance of sand.

It is impossible to give precise directions for drawing facies maps or to say exactly what zones should be distinguished in every case or what stratigraphic units or volumes should be shown on one diagram. If the sequence consists of numerous alternations of rock types, the diagram may show an alternation of clays and sands, of clays and limestones with a predominance of the former, of stratifications of clays with individual interbeds of limestones, etc. If there is a considerable change in facies in the vertical direction within a subdivision that has been treated as a single unit on the chart of the thicknesses, this unit must be divided up and a more detailed facies diagram drawn. The number of such subdivisions according to facies will be decided by the part of the section where the facies are most complex: for example, if at point A the entire Devonian is made up of clays, whereas at point B the Lower Devonian is composed of sandstone, the Middle Devonian of clay, and the Upper Devonian of limestone, one must draw three facies maps on which point A will appear the same and the differences will be seen only at point B.

In constructing facies maps one must pay special attention to the exact temporal relationships of the deposits shown on them. If Lower Viséan sediments appear in one part of a map of the Viséan stage and Upper Viséan facies on another part of the same map, the result will be meaningless. Finally, in drawing facies maps one must resort to interpolation, which will at times be fairly extensive.

The distribution of clastic sediments of various grain sizes on the map will point to the locations of areas of erosion, which are indicated by a change from zones of finer to zones of coarser deposits. Sometimes the material makes it possible to determine the boundaries of erosional areas quite accurately.

One frequent misunderstanding must be pointed out. It is often assumed that as one approaches the source of the sediments, not only do the fragments become larger but also the thicknesses increase. Two different phenomena are being confused here. If the sediments are composed of both clastic and biogenic rock, the amount of terrigenous rock will increase in the direction of the area of erosion and the relative thickness of the clastic facies will thus also increase. But the thickness of the entire stratigraphic unit may nevertheless decrease, to the extent that it depends only upon the subsidence of the earth's crust in the region of sedimentary accumulation. At the border between the areas of erosion and sedimentation, of course, the thickness will diminish to zero; this decrease in thickness may begin far from the border and continue gradually, or there may be a zone of great thicknesses in the vicinity of the area of erosion that wedges out rapidly to zero as one enters the area itself. All these variations are of no significance in determining the location of the dry lands, since they depend entirely on the distribution of the oscillatory movements, which determine the

thicknesses of any sedimentary deposits. The locations of sources of sedimentary material can be found only on the basis of the changes in facies.

Diagrams of the facies and thicknesses may be combined on one chart, although this is not always suitable.

By drawing facies maps for successive stratigraphic units and comparing them, and also by preparing diagrams of the distributions of

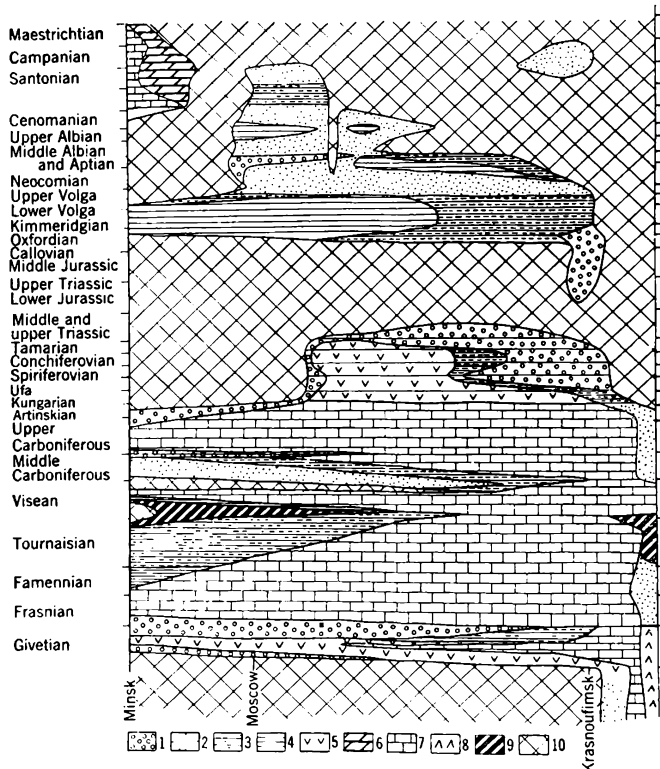


FIG. 106. Diagrammatic section showing facies through the Russian plain along the Minsk-Moscow-Krasnoufimsk line: (1) continental deposits; (2) sand; (3) sand and clays; (4) clays; (5) lagoonal sediments; (6) marls; (7) limestones; (8) volcanic rocks; (9) coal-bearing deposits; (10) areas of erosion.

thicknesses, one may determine the shifts in the erosional areas and the boundaries of such areas (and thus transgressions and regressions), the changes in facies zones with or without changes in their thickness, the changes in the distinctness of the boundaries between different facies zones (again in conjunction with changes in thickness), the migration of facies in connection with movements of the shore lines, and changes in the extent of the facies zones, etc. The results of such analyses will produce important conclusions regarding the oscillatory movements of the crust of the earth.

It was said above that the facies may be plotted on a profile of the thicknesses. In some cases, however, it is more appropriate to construct a facies profile, with the geologic time scale (relative or absolute) rather than the thicknesses laid off on the ordinate. From such a diagram one can quickly see the time during which a given facies was formed, the gradations and changes in width of the facies zones, and all the areas of deposition, etc. Figure 106, for example, shows the periodic character of the changes in the areas of deposition and erosion on the Russian plain.

The same limitations must be applied to the facies method as to the thickness method. The greater the detail of the facies analysis, the greater the danger that local irregularities in the distribution of the facies will invalidate the general conclusions regarding the tectonic regime. On the other hand, a too general approach will overlook these same regularities. Only a careful examination of each case under consideration will determine the degree of detail that will be most effective in a given facies analysis.

STRATIGRAPHIC BREAKS

In addition to the thicknesses and facies, to reconstruct the history of the oscillatory movements of the earth's crust one must study the distribution in time and space of the evidences of erosion and stratigraphic breaks. These indicate the uplifts of the earth's crust.

In establishing the duration of an uplift, one can use the method of stratigraphic breaks to determine accurately only the moment at which the uplift ended and became subsidence, as shown by the accumulation of the next younger sedimentary deposits. This point in time is determined by the age of the beds lying immediately above the break. But the beginning of the uplift cannot be determined with such accuracy: the uplift may have begun at any time during the interval between the age of the youngest beds below the unconformity and that of the oldest beds lying above it. This question has already been considered above in the discussion of zero thicknesses, but a parallel study of the facies of the deposits involved may produce a more accurate solution to the problem.

Stratigraphic breaks are not ubiquitous. In the Caucasus, for example, the rocks of the Upper Cretaceous in some places grade evenly into the underlying rocks and in others show clear evidence of breaks and erosion.

One method is to construct a map of the contacts between various suites, showing areas where the transition is continuous from one suite to another and other areas where there are breaks (Fig. 107). The latter areas should also show the location and the horizon upon which

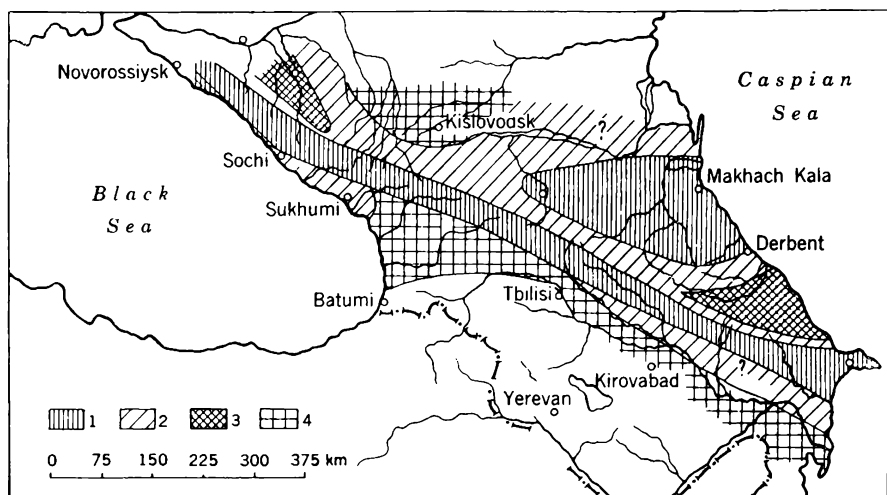


FIG. 107. Paleogeologic map of the Greater Caucasus during the pre-Callovian erosion: (1) areas of uninterrupted deposition of Callovian; (2) areas of transgressive deposition of Callovian sediments in the Middle Jurassic; (3) areas of transgressive deposition of Callovian sediments in the Lower Jurassic; (4) dry lands in Callovian time (after V. V. Belousov).

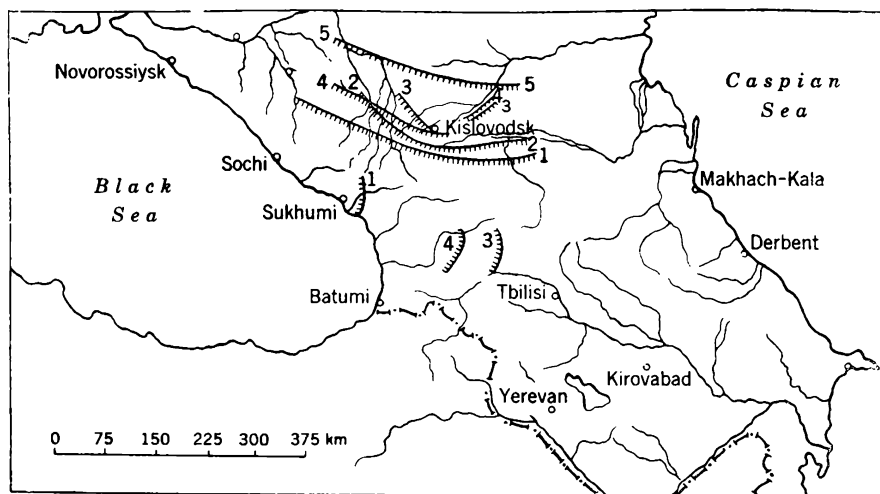


FIG. 108. Gradual development of the Upper Jurassic transgression in the Caucasus: (1-5) shore-line positions (1) in the Callovian age, (2) in the Lusitanian age, (3) in the Kimmeridgian age, (4) in the Tithonian age, and (5) in the Lower Cretaceous (after V. V. Belousov).

the bottom of the transgressive suite lies: thus the shape of the uplift that preceded the erosion and the new transgression will appear more distinctly. In certain regions of the Caucasus, for example, one may see that the base of the Upper Cretaceous lies first on the Albian, then on the Aptian, and still farther on the Neocomian stages, etc. Such maps

of the geologic structure of the surface of the unconformity are called *paleogeologic maps*; they are constructed not only to obtain theoretical information but also with the practical goal of prospecting for oil.

The most interesting cases are those in which the subsidence, after the erosional interval, develops and spreads gradually, successively involving greater areas. The result is that the basal horizon of the transgressive suite will be of different ages in different places. Maps showing the changes in the age of these horizons from place to place (Fig. 108) will present a picture of the gradual development of the transgression.

EXAMPLES OF THE TECTONIC INTERPRETATION OF VARIOUS SECTIONS THROUGH SEDIMENTARY ROCKS

G. F. Krashennikov (1945) described a simple and interesting case: in certain coal deposits the coal beds fork along the dip. The beds branch out, so to speak, so that a single bed occurs in one part of the deposit but in another is replaced by several thin beds separated by clastic rocks devoid of organic matter. Figure 109e shows a section of such a deposit. The succession of movements of the earth's crust that produced this result will be traced below.

The plant matter that forms the substance of coal accumulates in peat bogs in places where the water table rises to the surface of the earth. For a thick bed of coal to accumulate, the crust of the earth must subside at such a rate that the accumulation of plant matter will keep up with the tectonic subsidence. If the subsidence is too rapid, the peat bog will be flooded and become a lake or lagoon, whose bottom will be covered with deposits of clastic sediments.

The layer of coal is initially formed in the horizontal position. In the area shown at the left in Fig. 109 a single thick bed was formed; while this was being laid down, on the right side of the coal deposit coal beds I to IV and the clastic rock separating them were being formed.

In reconstructing the historical succession in the formation of these beds, one must remember that in the area shown in the right half of the illustration each bed of coal was originally formed in the horizontal position. To understand the manner in which the tectonic movement

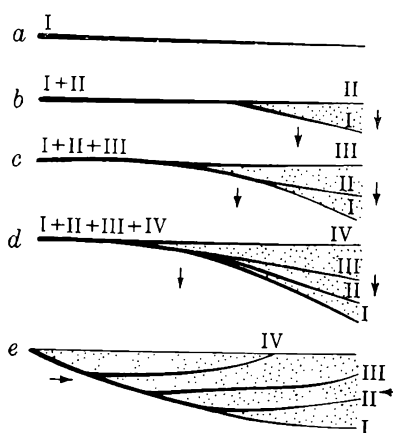


FIG. 109. Development of a coal deposit with branching coal beds: (a, b, c, d) successive stages in the development; (e) present-day structure (after G. F. Krashennikov).

developed, one must successively replace the beds, beginning with the bottom, in the horizontal position while maintaining the horizontality at the left (see Fig. 109*a, b, c, and d*).

For a certain length of time the subsidence encompassed the entire area with the same rate everywhere, so that only the single bed of coal I was formed throughout the whole area of the section (Fig. 109*a*). Then the subsidence became more rapid at the right, so that the speed of movement increased from left to right: the accumulation of peat could not keep pace with the subsidence, and part of the bog was flooded. At the bottom of the resulting body of water, above the peat, a stratum of clastic rock was deposited, at a rate equal to that of the subsidence, so that the thickness of the clastic material increased from left to right. At the right, meanwhile, the original slow subsidence was maintained, so that only peat accumulated.

At a certain point in time, however, the subsidence on the right side also slowed down temporarily to such a degree that the accumulation of plant matter could keep pace throughout the whole area of the deposit, and coal bed II was formed (Fig. 109*b*). This, of course, was deposited in the horizontal position. Below the clastic sediment separating bed I from bed II one can see that the former, as a result of the uneven subsidence of the earth's crust, now dips basinward.

Later the subsidence on the right side was accelerated again, resulting in a new accumulation of sedimentary rock, and again slowed down enough for coal bed III to be formed (Fig. 109*c*). At the left, meanwhile, the subsidence continued at the same constant slow rate. But since the boundary between the areas of different rates of subsidence was displaced to the left, and the area covered by rapid subsidence increased, the point at which bed III branches off from the collective bed is not the same as the point at which bed II branches off.

The following stage is the same as the preceding (Fig. 109*d*). After the deposition of bed III, the subsidence was again accelerated. The peat was again covered by clastic rock, except that the area of more rapid subsidence extended still farther to the left. Finally the entire region was again covered by a layer of peat (IV).

Figure 109*c* shows that the rapid subsidence eventually encompassed all the left side as well, for here also the coal bed, which up to this time had been continuous, is covered with clastic rock. Finally, all the beds, which were initially formed in the horizontal position and then gradually bent downward by the uneven subsidence, were subsequently uplifted on the right and subjected to erosion.

It is noteworthy that on the right side the lower coal beds are bent upward less than the upper beds. This results from the fact that the thickness of the clastic rock separating the coal beds increases from left to right. The final position of the beds is determined by the totality

of the preceding subsidence and succeeding uplift: the uplift involved all the beds equally, whereas the earlier subsidence had not been the same everywhere; thus the lower beds were downwarped and bent to a greater degree than those above, since they were subjected to the subsidence for a longer time and thus their final bending upward was correspondingly less.

All these conclusions have been based on a successive restoration of the individual basal horizons to the horizontal position. This procedure is of great importance in the tectonic analysis of a geologic section. Since in the case just cited the sedimentary accumulation kept pace almost exactly with the subsidence, the surface of deposition was always approximately horizontal. In the process of movement of the earth's crust beneath this horizontal surface, various thicknesses of sediment were deposited at various places, controlled by the amount of the subsidence. If, however, corrections must be made for a rate of accumulation that was not the same as the rate of subsidence, the surface of sedimentation must be represented not as being horizontal but as following the shape of the basin of sedimentation.

From this one can understand the sequence in the tectonic analysis of a section. First of all, the appropriate stratigraphic surfaces must be selected to mark the intervals: it is assumed that these are the surfaces dividing the section into systems, series, and stages. Each of these surfaces originally had the shape of the bottom of the basin, lower at points of greater depth of water and higher where the depths were less. If the original shape of any stratigraphic datum plane is restored and the thicknesses of the underlying strata are plotted below it, one will have a picture of the occurrence of the rocks at the time the basal horizon in question was formed. By proceeding in this manner from one surface to the next, one will obtain a picture of the entire history of the movements of this part of the earth's crust.

The history of the oscillatory movements of the Greater Caucasus will be considered as one more example, since it is contained in the distribution of thicknesses and facies in the profile in Fig. 105. In this case, corrections for the changes in facies will be disregarded, in view of the difference between the vertical and horizontal scales, and the stratigraphic horizons will be shown as straight lines. In the lowermost illustration the straight line represents the top of the Upper Jurassic; below it are plotted the thicknesses of the Upper Jurassic beds. Since the base of these strata was at one time also horizontal, it may be concluded from the shape that the movement of the earth's crust was uneven during Upper Jurassic time: the crust was more intensively downwarped in the area of the present-day ridge separating the Black Sea from the Caspian, and there was less subsidence in the area of the northern slope of the ridge and still less in the intermediate zone. At the same

time, within the area of the ridge there was uplift accompanied by erosion, as evidenced not only by the decrease in thickness northward and southward but also by the peculiar appearance of coarser facies in these directions.

In the next section the top of the Lower Cretaceous is in the horizontal position, and the thicknesses of this series and of the Upper Jurassic are plotted below it. The datum planes now have the shape they had at the end of the Lower Cretaceous. It is evident that the tectonic movements continued to be uneven but followed the same pattern as earlier: the zones of greater and less subsidence are in the same positions as in Upper Jurassic time. The Lower Cretaceous sediments were deposited north and south of the Caucasus Range, in places where there are no Jurassic sediments, either because they were not deposited or because they were eroded during the Jurassic period. Thus, in the Lower Cretaceous, the whole region of subsidence and sedimentary accumulation was extended.

The next profiles, for the Upper Cretaceous, the Eocene, and the Oligocene, show that the distribution of greater and less degrees of subsidence and the resulting accumulation of varying thicknesses of sediments remained approximately the same. In addition, the sequence from the base of the Upper Jurassic to the end of the Paleogene reflected the same differences in direction and rate of oscillatory movements as had been observed before.

A completely different picture is presented by distribution of thicknesses and facies in the Miocene and Pliocene. These suites, which are found only in the foothills, thicken northward and southward from the ridge, thinning again after a certain distance. The change in facies shows that at this time there was a great uplift with intensive erosion in the area of the ridge: the location of the greatest uplift coincided with the location of the greatest subsidence (in the southern zone) during the Mesozoic and Paleogene. During the Neogene there was subsidence only in the present-day foothills area.

In comparing these latter profiles with the earlier ones, it becomes obvious that at the boundary between the Paleogene and the Neogene there was a considerable change in the distribution of the oscillatory movements, particularly the change from subsidence to uplift in the center. The diagram of the present positions of the suites (Fig. 105, top section) summarizes all these movements. At the present time, for example, the downwarping of the Mesozoic deposits at the center of the ridge has changed to upward movement, but the amount of this upward movement in the lower beds is less than in the upper, since before the uplift the lower beds were subjected to greater subsidence than the upper. As regards the Neogene, the opposite situation obtained: the lower beds were uplifted to a greater extent than the upper, because

they were deposited at the same time as the gradual uplift of the ridge, and the older beds were upwarped more than the later beds.

Other examples of analysis of sections will be given below, along with an explanation of the laws governing the development of oscillatory movements and their history.

THE VOLUMETRIC METHOD

The volumetric method of studying the oscillatory movements of the earth's crust deserves special attention; this was developed and applied by A. B. Ronov (1949) to a number of cases.

The essence of this method is as follows: Any sufficiently large territory is divided, for any length of geologic time, into areas of uplift and of subsidence of the earth's crust; these movements will occur at different rates of speed in different places. The *volumetric* rate of uplift or subsidence is the rate of growth of the total volume of all the tectonic upwarps and downwarps of the earth's crust in the given territory; this net volumetric rate of movement of the earth's crust can be estimated quantitatively.

The volume of subsidence, as was seen above, may be considered approximately equal to the total volume of sediments deposited during the length of time in question (including extrusives). Uplift results in erosion; the amount of uplift is considered equal to the volume of clastic material that has been carried in, so that the total volume of the uplift may be determined by the volume of clastic rock. The problem thus is reduced to a determination of the volumes of all the accumulated deposits (including extrusives), on the one hand, and of the clastic deposits, on the other hand. The resulting figures will characterize the total volumes of subsidence and uplift during a definite length of geologic time. By following the changes in these magnitudes from one geologic time unit to another, one can obtain a quantitative expression of the changes in the overall intensity of uplift and subsidence. The mean rate of subsidence and uplift of the earth's crust is found by determining the volume of uplift and subsidence per unit area (one square kilometer) and of time (one million years).

The total volume of the rocks (U) is calculated from the thickness diagram. Accurate calculation is extremely difficult; for the first approximation, the procedure is as follows: the areas contained within each isopach are computed and a series of figures (a, b, c, d , etc.) are obtained, each pair of successive figures is added and divided by 2, the results are added again and multiplied by the magnitude of the section encompassed by the isopachs (X):

$$U = \left(\frac{a}{2} + b + c + \cdots + \frac{n}{2} \right) X$$

Much more complex operations are required to determine the volume of terrigenous rocks alone. In some cases, where large enough thicknesses are present, charts of the distribution of thicknesses of terrigenous suites can be made for this purpose. If there is an interbedding of biogenic and terrigenous rocks, one must establish the distribution of terrigenous rock for as many sections as possible as a percentage of the total volume of sediments under consideration, and from these figures draw isopachs showing the distribution of the total thicknesses of terrigenous interbeds.

Very frequently it is impossible to calculate either the total volume of rocks or the volume of clastic sediments, but the ratio between these volumes can be determined. In folded regions, for example, individual suites crop out on the surface in belts, and between these belts the suites are either eroded or hidden at depth. If a large number of variously shaped folds exists in the region in question, it may be assumed that the relationships of the outcrop areas of various suites on the surface is approximately equal to the ratio between their volumes. Thus the outcrop areas of terrigenous suites are measured on the map and compared with the total areas of the outcrops of the whole stratigraphic complex; this ratio, at first approximation, will be equal to the ratio of the volume of uplift and the volume of subsidence. By studying one stratigraphic subdivision after another in this manner, one can determine how this ratio changes with the course of time. In many cases, this is sufficient basis for important conclusions regarding the history of oscillatory movements.

One aspect remains to be elaborated. As long as the volume of terrigenous rock is compared with the total volume of all the rocks, the ratio can never be greater than unity. Consequently, this method cannot be used to demonstrate that the volume of uplift was greater than the volume of subsidence. Such cases do occur, however, and must be identified; to this end, corrections for changes in facies must be introduced. If the total uplift is greater than the subsidence, more material is brought in than can be accommodated in the free space made available by the subsidence and the surface of deposition will rise. The amount of this rise in the surface of deposition is the measure of the excess of uplift over subsidence. For purposes of calculation, the level that existed when the given complex of rocks began to be deposited (the initial depth of the basin) is taken as the initial level. Only the rocks below the initial level are included in the calculation of the total volume, whereas all the terrigenous sediments both above and below the initial level enter into the volume of terrigenous rock. The biogenic rocks lying above this level (such as limestones deposited in shallower water than those deposited initially) are never included in the calculations, since they correspond neither to any subsidence of the earth's crust

equivalent to their thickness nor to the uplift in adjoining areas. It will be assumed that the first beds of the series in question were deposited at a depth of 200 m and that the section ends with sediments deposited at a depth of 100 m. In calculating the total volume of rock, the uppermost 100 m of thickness is excluded, but in calculating the volume of terrigenous rock all the clastic rock in any part of the section is taken into account. Corrections for the changes in facies can easily be introduced in cases where the surface of deposition was lowered (the basin became deeper) during the time covered by the investigation. Then all the volume of space below the initial level must be included in the

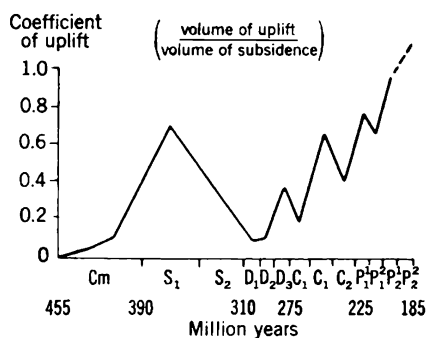


FIG. 110. Quantitative expression of the change in the ratio of uplift to subsidence in the Urals during the Paleozoic (*after A. B. Ronov*).

total volume of rock, even though it is only partially filled with sediments.

Figure 110 shows the results of calculations of the ratio of uplift to subsidence in the Urals from the Silurian through the Permian periods inclusive. It can be seen that this ratio changes from one epoch to another.

In such calculations, it is of great importance to determine the boundaries of the region for which the study is made. Since volumes of clastic rock are being studied, all areas within this region that are underlain by terrigenous material are automatically included. This terrigenous material, however, might have originated from various sources lying in different directions from the area of deposition and having areas of different sizes. Thus, to avoid serious errors, the calculations must encompass extremely large areas.

CHAPTER 15

GENERAL FEATURES OF THE OSCILLATORY MOVEMENTS OF THE EARTH'S CRUST

The methods set forth above can be used to study the nature and history of the oscillatory movements of the earth's crust over great areas and long periods of time. It must be mentioned as a limitation, however, that all conclusions have been based on geologic sections through the surfaces of present-day continents. The ocean bottoms cannot be directly studied, and thus the greater part of the earth's surface cannot be used in deriving geotectonic laws. But indirect evidence shows with a great degree of probability that the tectonic processes in ocean basins are no different in principle from those on the continents (see Chap. 31).

VARIETIES OF OSCILLATORY MOVEMENTS

The history of the oscillatory movements of the earth's crust shows that two basic varieties of such movements may be distinguished.

1. Movements that may be called *oscillatory*. They take the form of a long-lasting and slow development of areas of uplift and subsidence in the earth's crust, which maintain their positions during a given period of time but are changed in a regular manner by wavelike movements along the earth's surface. It will be seen that these movements in geosynclines differ from those on platforms.

2. Movements of the earth's crust that may be called *general oscillations*. These movements usually take place at one time over large areas that include both geosynclines and platforms equally. They are divided into many orders. Typical among them are episodic and transitory movements that take the form of a rapid uplift and renewed subsidence of a large portion of the earth's crust, as a rule reflected by gaps in the stratigraphic sequence. This category of oscillatory movements includes slow, long-lasting pulsations of the earth's crust, which determine the

fundamental rhythm of the oscillatory movements of various orders (see below).

It is not impossible that a deeper study of both varieties of oscillatory movements will ultimately indicate that they are dissimilar tectonic phenomena. Perhaps the movements that have caused the formation of the continents and the ocean basins should be distinguished as a special category of oscillatory movements.

At the present time general oscillations have been little studied, whereas a number of laws governing the development of wavelike movements have been established. Henceforth the discussion will be concerned mainly with wavelike movements.

The existence of varieties of oscillatory movements was noted earlier by the present author (Belousov, 1938–1940). The question of dividing oscillatory movements into two categories was posed very clearly by V. Ye. Khain (1939), who called the undulatory movements *epeirogenic* and the general oscillations *oscillatory*.

Ubiquity and constance are the chief features of both varieties of oscillatory movement. Traces of oscillatory movements are found in geologic sections of any locality and in strata of any age. The oscillations are reflected in bedding, facies changes, indications of transgressions and regressions, deposits of different thicknesses, and stratigraphic gaps. The conclusion drawn from this is that the earth's crust has never and nowhere ceased to oscillate and never remains at rest.

REVERSIBILITY OF OSCILLATORY MOVEMENTS

One property of these movements is their reversibility, that is, the fact that movements in different directions in the same locality replace each other during the course of time. This is true of both wavelike movements and general oscillations. The area of the present-day uplifted Caucasus ridge was the site of predominant subsidence during the Mesozoic and Paleogene. In the area of the middle and lower Volga the earth's crust subsided from the beginning of the second half of the Middle Devonian to the beginning of the Triassic, when it began to be uplifted. The subsidence was renewed in the Lower Jurassic and again changed to uplift in Tertiary times.

Reversed oscillations may also be considerably smaller in amplitude. One example is the stratigraphic section of the northern Caucasus region south of Maykop. The accumulation of Triassic deposits (subsidence) was followed by erosion (uplift). Later on there was an accumulation of Middle and Lower Jurassic sediments (subsidence). Then, before the Upper Jurassic, there was another period of erosion (uplift). Sediments of the Upper Jurassic were then deposited, but the uppermost layer of these is separated from the underlying ones by traces of erosion (subsidence).

ence and uplift). A number of stratigraphic gaps appear in the Cretaceous and Paleogene deposits, where one can see the evidence of repeated alternations of subsidence and uplift.

But the uplift following subsidence, or the subsidence following uplift, are usually not the same as the preceding movement in amplitude or in the area encompassed. They are not mirror images of each other, and the usual transition from subsidence to uplift, or vice versa, is accompanied by some degree of redistribution of the movements. Thus the oscillatory movements are not a simple fluctuation about a single point. They are a complex process that constantly changes with time.

SMALL INTERRUPTIONS IN GENERAL OSCILLATIONS

It has been said above that the change in the lithology of the rocks in the vertical direction is, as a rule, not gradual but in stages, as determined by the bedding. This shows that the relationship between subsidences and uplifts also changes rather sharply. The surfaces of division between beds usually testify to a temporary cessation in the process of subsidence of the earth's crust. On this basis, one may speak of the existence of small interruptions in the oscillatory movements.

This conclusion does not contradict what has been said above regarding the continuous development of these movements. These two cases involve phenomena of different orders. Oscillatory movement is continuous in the sense of there being a constant upward or downward movement. But each displacement of the earth's crust is temporarily interrupted by small oscillations in the opposite direction. The moments at which the movement changes direction are pauses in the scale of time, measured by the continuity in the formation of the bed and thus relatively very short. The small interruptions are caused by the pulsating nature of the general oscillations.

RHYTHM OF GENERAL OSCILLATIONS

An important property of oscillatory movements is their rhythm. Although areas of uplift and of subsidence exist simultaneously on the surface of the earth, for the most part either upward or downward movement predominates over movement in the opposite direction and this predominance periodically changes from one direction to the other. These periodic changes are reflected in the periodic facies changes in the section.

On the Russian plain there is a predominance of terrigenous sediments at the base of the sedimentary strata of the Lower Paleozoic and of the Middle Devonian beds. The Upper Devonian and Carboniferous are represented chiefly by limestones and the Permian system again by

terrigenous as well as lagoonal deposits. The Paleozoic section ends with the red-colored continental Tatarian suite. The Jurassic beds lying above the Tatarian, after an erosional interval, are made up of terrigenous rocks. The Lower Cretaceous has the same composition. The Upper Cretaceous is represented primarily by limestones and marls, but clastic rocks are found again in the Tertiary system. In Quaternary time there is erosion.

From this succession of sediments one can see that at the beginning (in the Middle Devonian) uplift was very important and produced a great amount of clastic material. Subsidence was predominant in the Middle Paleozoic; during this time slight intermittent uplifts caused only a small amount of clastic material to be accumulated in the basin. Uplift predominated again at the end of the Paleozoic era, and the entire territory was finally raised above sea level. The facies of the Mesozoic and Cenozoic testify to the recurrence of another such complete cycle, with transition from uplift to subsidence and back again (uplift predominated in the Jurassic and Lower Cretaceous, subsidence in the Upper Cretaceous, and renewed uplift again in the Cenozoic). This change in facies can be understood from what has been said in previous chapters.

Similar observations of changing facies and of transgressions, regressions, and gaps in the depositional sequence in other regions lead to the same conclusion: in the development of oscillatory movements the overall predominance is sometimes one of uplift and sometimes one of subsidence.

As far as can be judged, these changes in the direction and nature of oscillatory movements take place with a certain *regular periodicity*. This takes the form of very large cycles, with periods of about 150 million years in the history of the earth. These are the great tectonic cycles, or stages of the first order, of which the last three, distinguished as the Caledonian, Hercynian, and Alpine cycles, are more or less well known. These three cycles encompass a great part of the earth's surface, including Europe, America, part of Asia, and certain other regions. The Caledonian cycle begins in the Cambrian or at the very end of the Proterozoic and ends at the beginning of the Devonian. The Hercynian cycle begins in the Devonian and ends in the Permian period. The Alpine cycle occupies the whole of the Mesozoic and Cenozoic eras.

At the beginning of each cycle, uplift predominates to a great degree over subsidence, the areas of subsidence are small, and the areas of uplift are extensive and produce a great amount of clastic material. The predominance then gradually passes to the downward movements, which prevail at the middle of the cycle: the areas of subsidence increase in extent, and the subsidence itself becomes more intense. Toward the end of the cycle uplift again predominates; thereafter the next cycle begins.

Actual cycles will be examined on the basis of factual material in the next chapter. The approximate coincidence in time of the changes in the direction of the oscillatory movements in various parts of the earth's surface is one of the most striking and regular features of the earth's history. It must be mentioned, however, that in the case of the last cycle—the Alpine—there is an essential discrepancy between the areas encircling the Pacific Ocean, on the one hand, and the remainder of the surface of the continents on the other, which may for these purposes be called the *Atlantic area*. In Europe and the Atlantic part of North America the maximum subsidence in this cycle is in the Upper Cretaceous and the uplift at the end of the Tertiary period, whereas in East Asia, for example, uplift is already predominant at the beginning of the Cretaceous period. For this reason the concepts of the Caledonian, Hercynian, and Alpine cycles, as they are usually understood, are strictly applicable only to the Atlantic zone and must evidently be changed for the Pacific area. Moreover, in speaking of these cycles, one has in mind their typical appearance in Europe.

The great cycles are the basic landmarks in the history of the earth's structure. There have been many such cycles; if their lengths are the same and if the age of the earth is taken to be 4 billion years, there must have been about twenty-five cycles during the earth's past history. Of these, only the three latest can be studied, because the history of the earlier cycles is hidden in the poorly differentiated strata of the Precambrian.

Numerous present-day writers divide the Alpine cycle into two parts: an Alpine phase proper, which took place only during the Cenozoic, and a Mesozoic phase (Pacific, Nevadian, Yen Shan, Cimmerian). In this case four cycles are distinguished as having taken place since the beginning of the Paleozoic, and not three, as in this volume. The present author does not distinguish the Mesozoic as an independent cycle of the first order: he considers that in the Pacific Ocean area the same Alpine cycle ran its course as in Europe but that its duration was displaced in time.

The approximate correspondence in time of great cycles over large areas, and the almost completely simultaneous and ubiquitous increases in the rates of subsidence and uplift, are the reasons for the simultaneous occurrence of the great marine transgressions and regressions.

Historical geology has shown that the periods of very widespread transgressions, that is, of expansions of the sea and contractions in the areas of dry land, are the following:

1. The Cambrian period, corresponding to the first half of the Caledonian cycle.
2. The second half of the Devonian and beginning of the Carboniferous periods (the first half of the Hercynian cycle).

3. The Jurassic and Cretaceous periods (the first half of the Alpine cycle).

Regressions, that is, contractions of the seas and expansions of the dry land areas, were most widespread in the following periods:

1. The Silurian period (the second half of the Caledonian cycle).
2. The end of the Carboniferous and the Permian periods (the second half of the Hercynian cycle).
3. The Tertiary and Quaternary periods (the second half of the Alpine cycle).

Figure 111 shows an overall chart of the movements of the earth's crust. This chart depicts the history of the movements in the Atlantic area of the earth.

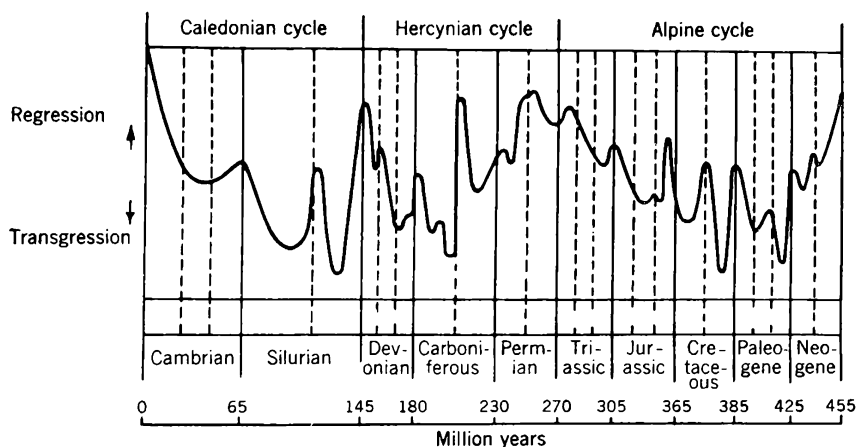


FIG. 111. Diagram of the oscillatory movements of the earth's crust.

The simultaneous occurrence of the great undulations of the earth's crust has been established for the continents. The movements that took place during these times in the oceans are not known. But if these movements had developed in the same direction and with the same amplitude as on the continents—that is, if the cycles were produced by an overall uniform expansion and contraction of the earth over enormous areas—it would have been impossible to observe the great transgressions and regressions that have taken place on the continents: such overall contraction and expansion of the volume of the earth and its surface would have led to negligible increases and decreases in the mean depth of the oceans. The sea level could have fluctuated only within the limits of a few meters. One must therefore conclude that oscillations of the earth's crust take place in the oceans as well but that these are either in the direction opposite to the movements of the land masses or else in the same direction as on the continents but with a small amplitude. Moreover, during periods of uplift, the relative height

of the continents over the ocean bottoms increases and during periods of subsidence, in turn, decreases. At times there is an accentuation and at other times a leveling off of the largest and most prominent features of the topography; these very changes in the relative height of the continents are, fundamentally, due to the transgressions and regressions.

It may thus be said that the periodicity of the oscillatory movements results from the superimposition upon the wavelike movements, which produce uplifts and subsidence, of general oscillations that take the form of great and long-lasting pulsations of large parts of the earth's crust.

COMPLEXITY OF GENERAL OSCILLATIONS

Smaller cycles, of the second, third, or higher orders, are always superimposed on the great cycles as a complicating factor. The relationships between these cycles must be understood as a superposition of pulsations of different sizes upon each other (see Fig. 111). This causes the complexity of the oscillatory movements; that is, the subordination of smaller oscillations to greater ones. Small cycles, like large ones, are reflected in the alternation of transgressive and regressive series of deposits and in the advances and retreats of the sea. But these small series occur within the larger series, and the changes in the facies and dimensions of the basins that produce them are smaller than those associated with the great cycles.

As an example, the Alpine cycle began with intensified subsidence and increased marine transgression, progressing from the Triassic through the Upper Cretaceous, inclusive. After this there was universal regression, ending with a great uplift of the continents at the end of the Tertiary and during the Quaternary periods. But, against the major transgression of the first half of the cycle, there were partial regressions at the ends of the Jurassic and Cretaceous periods; these regressions were temporary and rapidly changed to new transgressions. It may thus be said that the overall transgressive direction of the development was interrupted for a brief time and then renewed. In many places, moreover, the Lower Cretaceous (often together with Tithonian) as well as Paleogene deposits overlie an eroded surface of the deposits of previous periods. This is the evidence left by second-order cycles, whose duration corresponded approximately to that of a geologic period: 30 to 40 million years.

These partial cycles do not encompass so great a territory as the first-order cycles. They take different forms in different regions, and the higher the order and the smaller the duration of the cycle, the greater the variety in the development of individual parts of the earth's crust during the given cycle.

In Europe, the following second-order movements have been established as having taken place against the background of the basic cycles:

1. In the Caledonian cycle, (*a*) a regression at the end of the Cambrian and (*b*) a regression in the middle of the Silurian (at the boundary between the Ordovician and the Gotlandian).

According to these regressions, the great Caledonian cycle is divided into three second-order cycles: the Cambrian, the Lower Silurian, and the Upper Silurian. Each of these, like the great cycle, begins with an intensification of subsidence and ends with an increase in uplift. They are distinguished from the basic cycle by the fact that the uplifts that conclude the second-order cycles are never as extensive or as lasting as the uplift at the end of the first-order cycle.

2. In the Hercynian cycle, a regression at the middle of the Carboniferous period. This divides the greater Hercynian cycle into two second-order cycles.

3. In the Alpine cycle, (*a*) a regression at the end of the Jurassic period, (*b*) a regression at the end of the Lower Cretaceous, (*c*) a regression at the end of the Upper Cretaceous, and (*d*) a regression at the end of the Paleogene.

The Alpine cycle may thus be divided into five smaller cycles: the Kimmeridgian (Triassic-Jurassic), the Lower Cretaceous, the Laramide (Upper Cretaceous), the Paleogene, and the Neogene.

Secondary cycles of movement, in the localities where they appear, in one way or another affect the facies of the sediments or the boundaries of marine basins. The directly observable transgressions and regressions that correspond to partial cycles are always the result of a cumulative combination of second- and first-order oscillations. Thus, for example, the Upper Cretaceous transgression is distinguished for its great extent and wide distribution: it is the result of a second-order subsidence and of the great subsidence that marks the middle of the great Alpine cycle. In contrast to this, the Paleogene transgression nowhere attained such great extent, since in this case the secondary subsidence was superimposed on the upward-moving part of the great cycle.

Secondary cycles, in turn, differ in different localities, since they are complicated by smaller cycles, which leave traces in the form of smaller gaps in the stratigraphic column, and the latter, in turn, by still smaller ones, down to the very briefest cycles, which run their courses during the formation of a single bed. The formation of the bed and of the surface that separates it from the succeeding bed is evidence of a periodic change from subsidence of the earth's crust to standstill or to a certain amount of uplift, against a background of overall subsidence. The absolute duration of these minor cycles in years cannot be determined, because of the inadequacy of the available methods. Certain investigators (Khain, 1939; Vassoyevich, 1940) have suggested that

there are cycles that last 4 to 6 million years (corresponding to a stratigraphic stage), several hundred thousand years (a horizon or zone), a few score thousand years, and 500 to 1,000 years.

Minor cycles sometimes develop over a limited territory with a surprisingly regular periodicity. The clearest example of this is the flysch suites, which are distinguished by their regular cyclic bedding, such as the Upper Cretaceous and Paleogene flysch series on the southern slope of the Greater Caucasus.

In this instance, a very thick series of rocks is divided into a large number of beds or groups of beds, each of which is composed of a definite "assembly" of rocks that follows a definite sequence. N. B. Vassoyevich (1951) has studied the Upper Cretaceous and Paleogene flysch of the southeastern Caucasus: here each full sequence or cycle begins at the bottom with a bed of coarser rocks (conglomerate, calcarenite, or fragmental limestone), followed by marl and clay above. The top layer of the sequence is usually eroded away, and the succeeding sequence again begins with a bed of the coarsest sediments. Each sequence is from 0.5 to 2 m thick. This rhythmic alternation of rocks of definite facies indicates a similar periodicity in the relationships between the rates of subsidence and uplift. These fluctuations, however, are superimposed on the background of a more general downward movement of the earth's crust that has produced a very thick accumulation of sediments.

Studies have shown that the flysch suite of the Caucasus also contains evidences of other movements that are somewhat more intensive and long-lasting than those which have produced the cyclic bedding. These latter movements have produced surfaces of erosion and stratigraphic gaps that are irregularly distributed within the series.

Thus it has been established that oscillatory movements of four orders have taken place in the Caucasus area, during the time in which the flysch series was deposited:

1. Movement on the scale of an entire cycle (such as the Alpine), taking the form of a major subsidence of the earth's crust.

2. Movement on the scale of a single period, or of a second-order cycle, such as the Laramide or Paleogene; during the deposition of the flysch series of the Upper Cretaceous and Paleogene, this movement took the form of uplift, manifested as erosion along the boundary between the Cretaceous and Tertiary systems.

3. Movements that have produced surfaces of erosion in the flysch series, which occur less frequently than those marking the change from one sequence to another.

4. Movements that have caused the formation of the cyclic bedding.

A more detailed study would doubtless reveal a greater number of orders of movements.

The totality of all the existing oscillatory movements of various orders may be called the *spectrum* of the oscillatory movements of the earth's crust.

The spectrum of the movements changes from place to place and from period to period. On the southern slope of the Greater Caucasus, for example, minor rhythms appear throughout the flysch zone of the Mesozoic deposits, whereas in moving to the northern slope of the Greater Caucasus one does not find these minor cycles in the series of the same age: the bedding ceases to be cyclic, and only the very much larger cycles can be discerned. On the other hand, on the same zone of the northern slope the spectrum changes with the transition from one suite to a series of another age.

Study of the spectra of oscillatory movements and their changes in space and time is an extremely interesting and very important task; nevertheless, almost nothing has so far been done in this direction. The complexity of the oscillatory movements and the superposition of oscillations of different orders and amplitudes on each other present special difficulties in determining the overall direction of movement of the earth's crust in a given concrete case, since movements in opposite directions superimposed on each other may be found to exist simultaneously. Against the background of a subsidence that is the movement of the first and greatest order for the area and time under consideration, for instance, there may be an uplift of smaller extent and of the second order, which, in turn, may be complicated by a still smaller subsidence, and so forth. The actually observable movement is thus always the algebraic sum of all these movements on different scales superimposed on each other. If a downward movement of the first order is more intensive than the subordinate uplift superimposed on it, the earth's crust will be lowered; with the opposite combination, it will be raised.

Resolution of the total oscillatory movement into its individual movements is possible only through a study of the history of these movements over a long period of time, so that one may discover and trace the undulations of different amplitudes and different durations.

It is evident from the above that it is the general oscillations that are meant in speaking of the complexity of oscillatory movements. Wavelike movements also are made up of several orders of movements, which take the form of large waves complicated by smaller waves of various dimensions.

DIFFERENT REGIMES IN THE DEVELOPMENT OF WAVELIKE MOVEMENTS

A study of the history of the wavelike movements over a great area during one of the latest first-order cycles (such as the Caledonian or

Alpine) will show that there are two types of areas in the earth's crust. In the first type these movements are characterized by great amplitude and a relatively high rate; in areas of the second type the movements are of comparatively small amplitude and low rates of development.

The Alpine cycle in the Greater Caucasus began with a very large subsidence of the earth's crust (Jurassic to Paleogene), which reached an amplitude of no less than 12 km, and then continued with a great uplift (Neogene to Quaternary) that led to the formation of the high mountain ridge. During this time the subsidence on the Russian plain was of small amplitude (600 m or less). The uplift here was also small.

The Hercynian cycle involved great subsidence and great uplift in the Urals and the Appalachians, while on the Russian plain and in the central part of the United States the oscillations during this time were small. During the Caledonian cycle there were very great oscillations of the earth's crust in the Urals, in central Kazakhstan, in northern Tien Shan, in the Altay, and in Scandinavia. At the time of these movements, however, the central states of the U.S.A., a great part of South America, and other territories were in a state of comparative rest.

One characteristic feature of the more mobile regions is that the major subsidence in them changes with the course of time to a major uplift and that there are, simultaneously, adjacent zones that differ sharply in the direction of movement of the earth's crust. Directly adjoining the zones in which the great thicknesses of the deposited suites testify to the intensive subsidence of the earth's crust lie areas in which these same suites are either very thin, indicating a small amount of subsidence, or else are absent. The latter case is explained by the fact that in these areas there was uplift and erosion, the uplift, to judge by the clastic material brought in, being extremely rapid.

Since these zones that are so different in their behavior lie very close to each other, the thicknesses of the deposits change very rapidly and within short distances. If the change in thickness per unit distance is called the *thickness gradient*, the zones in which there is great mobility of the earth's crust may be characterized as regions of *large thickness gradients*, in accordance with the great speeds and amplitudes of the oscillatory movements and the great amount of contrast in them.

In distinction to this, regions of the second type, which are less mobile, are said to have *small thickness gradients*. These regions are also divided into areas characterized by different movements of the earth's crust, that is, by greater and less subsidence and uplift. But since the amplitudes of these movements are small, the transitions from greater to smaller or to zero thicknesses are slow and gradual and have small gradients. Here one is dealing with movements showing a small amount of contrast.

The difference in the gradients is clearly reflected by the distribution

of the isopachs. In mobile regions the isopachs are close together, and they spread far apart in the regions of small mobility. Figures 93 and 94 contrast the Greater Caucasus and the Russian plain (during the Mesozoic) in this respect. These diagrams clearly show the great magnitude of the thickness gradients in the Caucasus and their small magnitude in the Russian plain.

WAVELIKE MOVEMENTS IN GEOSYNCLINES AND ON PLATFORMS

Investigations have shown that in zones in which the oscillatory (wavelike) movements are of considerable magnitude and great contrast, the other varieties of tectonic movement are also distinguished by their great violence. Such zones are the sites of intense folding and faulting movements and of varied and extremely active igneous processes mainly in the form of major intrusions.

These cases thus involve portions of the earth's crust that are characterized by generally high mobility and by considerable changes in their structure. As a result of these changes, the mobile region is transformed into a folded zone with the complex structure that has already been remarked upon. Such areas of particularly great tectonic mobility, as is known, are called *geosynclines*. This term was suggested by the American geologist Dana (1873), who believed that the great thicknesses (in comparison with other areas) of the sedimentary strata in such zones testify to a great subsidence of the earth's crust and to the formation of an enormous synclinal flexure. At the present time, more complex ideas have been evolved regarding the structure and history of mobile zones. But in spite of the more complicated nature of these concepts, it appears that one of the most important processes in the history of a geosyncline is the great subsidence of its individual parts, so that the old term may quite properly continue in use.

A few words must be said about the difference between the meanings of the terms "geosyncline" and "folded zone." *Geosyncline*, in the author's understanding, is a historical concept that refers to a definite combination and succession of intensive tectonic movements that take place in a certain area. The term *folded zone* is a structural and morphologic one, which designates the complex structure that appears as a result of the geosynclinal development of a certain region.

In many cases it is very useful to classify tectonic terms as historical or dynamic, on the one hand, and structural or morphologic, on the other.

Those regions in which the wavelike oscillatory movements are of small magnitude and contrast, in distinction to geosynclines, are characterized by the weak development of other tectonic movements as well. In these areas folding movements are extremely poorly developed, and

intrusive igneous activity is almost completely lacking. Such regions are called *platforms*. This term is used by the author both in a historical sense, to reflect the complex of gentle tectonic movements typical of such areas, and in a structural sense, referring to the structure of platforms and the specific forms in which the rocks in them occur.

Thus the oscillatory movements in geosynclines are characterized by great magnitude and contrast and by relatively rapid movements of large amplitude. On platforms these same movements are of small magnitude and contrast and have low rates of movement and small amplitudes.

Inasmuch as oscillatory movements are, in the author's opinion, the fundamental forms of tectonic movement, their mode of development (great or small contrast and great or small magnitude) is an essential part of the concept of geosynclines and platforms. But a complete definition of geosynclines and platforms, besides oscillatory movements, must include the mode of development of folding and faulting movements and igneous activity.

It would be desirable to support the qualitative differences between the oscillatory movements in geosynclines and platforms by quantitative data: what are the thickness gradients, the magnitudes of the movements, and the rates of speed that characterize a geosyncline and a platform? It is exceedingly difficult, however, to provide such quantitative definitions. Measurements of thicknesses show that strata as much as 12 to 15 km thick are accumulated in areas of the greatest subsidence in geosynclines. The subsidence must therefore have actually been greater, since, as will be seen below, the thicknesses of the deposits in folded zones are seen not in their primary form but only after they have been compressed by mechanical forces. Allowing for this decrease in the thickness, it may be thought that the maximum subsidence of the earth's crust in geosynclines may attain 20 to 25 km. These figures indicate the approximate magnitude of the oscillatory movements of the earth's crust. They are negligible in comparison with the radius of the planet, but in relation to the thickness of the earth's crust (if this is taken to be 40 km) they are quite considerable.

The rate of subsidence of the earth's crust in geosynclines, according to average data, is 0.1 to 0.2 mm per year. This rate is lower than the one mentioned above for present-day oscillatory movements. The reason for this discrepancy is that the rate of movement for earlier periods is calculated by dividing the thickness by the absolute length of the period in years and thus obtaining the average rate of the total subsidence over a great length of time. The overall movement, however, is always complex and made up of many upward and downward oscillations. One can easily see that the rate of these partial movements must be considerably greater than the average rate that results from the

totality of the movements. But in studying present-day movements one is dealing only with these subsidiary movements. The magnitude and rate of the movements on platforms are much smaller than in geosynclines. Since there are intermediate types of movement, it is hard to draw accurate boundaries between platforms and geosynclines; nevertheless, for the most typical cases, the figures for platforms are approximately ten to twenty times less than for geosynclines.

The contrasting wavelike oscillatory movements in geosynclines divide them into parts with great subsidence of the earth's crust, on the one hand, and other parts with less subsidence or with uplift, on the other. Ultimately, disregarding the smaller absolute subsidence and keeping in mind only the relative uplift that always appears in areas of the second type, these latter may for the sake of brevity be called *zones of uplift*.

The subsidence and uplift that are always found in geosynclines play an extremely important role in their development and structure. Up to the present time, however, no single set of terms has been evolved to designate the areas of subsidence and uplift within the geosyncline.

A. D. Arkhangel'skiy (1937, 1939), N. S. Shatskiy (1946), and a number of other investigators refer to internal geosynclinal subsidences and uplifts as geosynclines and geanticlines, respectively, and refer to the entire mobile zone as a whole, including a number of subsidences and uplifts, as a *geosynclinal region*, or *geosynclinal system*. Others speak of synclinoria and anticlinoria, of depressions and upwarps, or of block mountains, anticlinal uplifts, and synclinal subsidences. The author of this book in his writings uses the terms *internal geosynclines* and *internal geanticlines*, or *intra geosynclines* and *intra geanticlines*, reserving the term *geosyncline* for the entire mobile zone as a whole. For the final stage of development of a geosyncline, however, *uplift* and *subsidence* and several varieties of these terms have been used.

V. Ye. Khain (1946), having in mind the existence of subsidences and uplifts of various orders within mobile zones, calls intra geosynclines and intra geanticlines *minor subsidences* and *uplifts*, geosynclines and geanticlines *major subsidences* and *uplifts*, and the entire mobile zone a *geosynclinal region*.

Here the terminology adopted earlier by the author will be used; that is, the entire mobile zone as a whole will be spoken of as a geosyncline, and within it intra geosynclines, intra geanticlines, uplifts, and subsidences will be distinguished. The differences between intra geosynclines and intra geanticlines, on the one hand, and uplifts and subsidence, on the other, lie in the time of their formations. These differences will become clear from what follows.

Certain remarks are in order here in regard to certain attempts to

bring more precision into the terminology of geosynclines. In the author's opinion, it is incorrect to use "geosyncline" for a zone of subsidence that forms part of a greater mobile region. This term should be reserved for the whole *mobile zone*, to indicate its unitary nature. When the term is used otherwise, the impression is unavoidably created that the geosynclinal region is composed of a number of completely independent geosynclines having no connection with each other and that the word thus has a purely geographic meaning.

It is also incorrect to use "synclinatorium" and "anticlinorium" to designate geosynclinal subsidences and uplifts. These words should be used in a purely *structural sense* only. They must be associated with the term "folded zone" and not with "geosyncline," which has a historical and dynamic meaning.

Anticlinal uplift and *synclinal downwarp* are all the more inappropriate terms because the words "anticlinal" and "synclinal" are, in this case, superfluous.

Geosynclines, intrageosynclines, and intrageanticlines will henceforth, for the sake of brevity, often be designated as GS, IGS, and IGA, respectively.

What distinguished the geologic section of an intrageosyncline from that of an intrageanticline? Figure 112 shows three stratigraphic columns of the Lower Cretaceous in the northern Caucasus. The left column shows the intrageosyncline of the main Caucasus ridge (Pshekha River), the center column the intrageanticline of the northwestern Caucasus (Belaya River), and the right-hand column the northern Caucasus intrageosyncline (Assa River).

The intrageosynclinal sections, naturally, contain great thicknesses, which correspond to the great subsidence of the earth's crust. The thickness of the Lower Cretaceous deposits on the Pshekha River is 2,200 m and on the Assa River 1,300 m, whereas on the Belaya River it is only 300 m or less. The first and third columns also show a more complete stratigraphic section, without gaps in the accumulation of sediments that would cause whole formations to be missing from the section. There are several gaps in the intrageanticlinal section (Belaya River), and whole stages—the Valanginian and the Albian—are missing from the section.

The difference is easy to understand. The vertical movement of the earth's crust always takes the form of a zigzag, and the subsidence is continually being interrupted by upward movements. In the intrageosynclines subsidence predominates to a great degree over uplifts, which are of relatively small importance and merely cause a certain redistribution of the material and the formation of minor, frequently insignificant gaps in the sequence. Within the intrageanticlines, since the net subsidence proceeds slowly, movement in the opposite direction plays a large role and many times produces a long-lasting uplift of the surface of deposition above sea level, thus leading to erosion.

Although the entire geosyncline subsides, the full extent of this subsidence is seen only in the intrageosynclines. In other areas (the intrageanticlines), local uplift is superimposed as a second-order phenomenon upon the overall subsidence. If the former is greater than the general subsidence, it will result in an absolute uplift of the earth's crust; if it is less, it will merely cause a slower subsidence in the intrageanticlines than in the intrageosynclines. The intensity of the overall subsidence and the internal uplifts changes with time and in space; from this

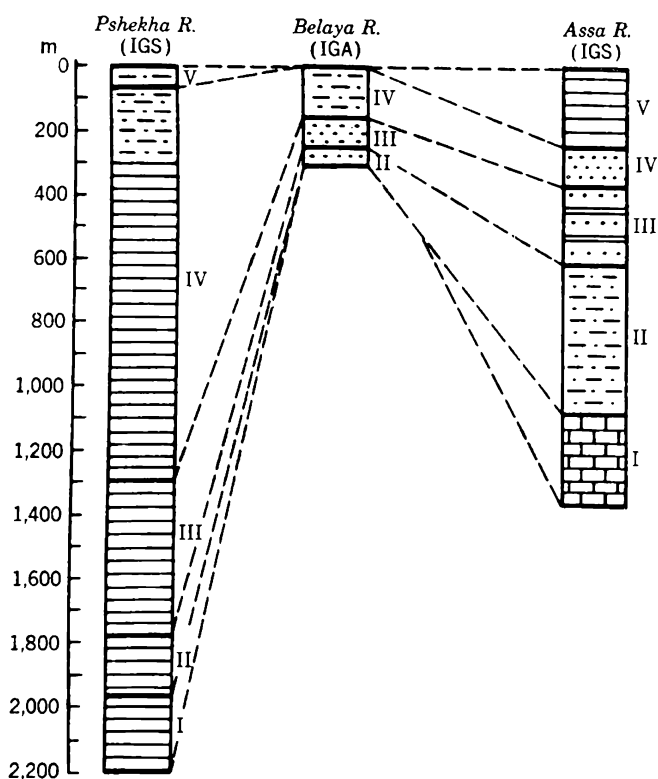


FIG. 112. Three stratigraphic columns of Lower Cretaceous deposits in the northern Caucasus (IGS-IGA-IGS): (I) Valanginian; (II) Hauterivian; (III) Barremian; (IV) Aptian; (V) Albian. For meaning of abbreviations, see text.

it is clear that the conditions producing a total lack of deposition in the intrageanticline and the conditions leading to the accumulation of small thicknesses of sediments are related and must grade into one another.

Platforms are also divided into areas of relatively greater subsidence and areas of less subsidence or of uplift. In the case of platforms there is also no single terminology for these features. The subsidences and uplifts formed by the oscillatory movements on platforms are called *basins* and *arches*, *mobile* and *stable shelves*, *depressions* and *uplifts*, *hollows*

and ridges, *synclises* and *antecclises*, and even *geosynclines* and *geanticlines*.

The subsidences and uplifts on platforms will here be called *subgeosynclines* and *subgeanticlines*; these terms will be used in the historical, dynamic sense. They will serve to underline the partial similarity between subgeosynclines and subgeanticlines, on the one hand, and intrageosynclines and intrageanticlines, on the other. In referring to the structure, the terms *syncline* and *antecline*, already introduced above, will be used. For brevity, subgeosynclines and subgeanticlines will be designated as SGS and SGA, respectively.

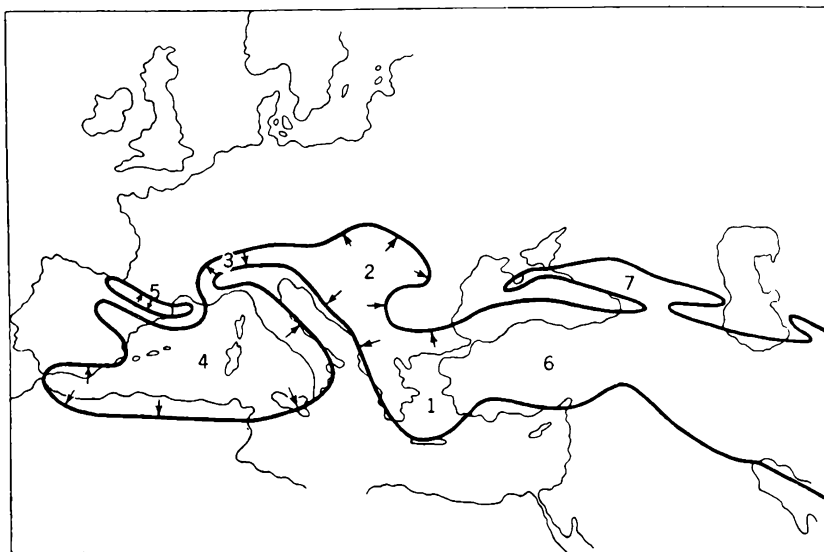


FIG. 113. Map of the Alpine geosyncline in the Mediterranean region: (1) Aegean oval; (2) Hungarian oval; (3) Alps; (4) western Mediterranean oval; (5) Pyrenees oval; (6) Asia Minor oval; (7) Caucasian oval.

The same differences, generally speaking, exist between subgeosynclines and subgeanticlines as between intrageosynclines and intrageanticlines. In the first of each pair the thickness of the deposits is greater and the section is stratigraphically more complete than in the second, in which deposition may be totally lacking.

In regard to the problem of the sizes and shapes of platforms and geosynclines, the idea is widely held at the present time that geosynclines form long, extended belts that stretch out between the more rounded platforms. This concept requires more precise definition.

Figure 113 shows a diagram of the Alpine geosyncline (that is, the Alpine cycle) in the Mediterranean region. The areas of more active oscillatory movement are outlined by contour lines, which reveal the

complexity of the geosyncline's shape. The geosyncline is composed of separate, for the most part unenclosed, ovals that adjoin each other. Among these are the western Mediterranean, the Hungarian, the Aegean, and the Caucasian ovals. The Alps form a narrow bridge between the first and second ovals. One interesting feature is the geosynclinal region of the Pyrenees, which is self-contained and not connected to the main body of the geosyncline. It should also be noted that the Alpine geosyncline of the Mediterranean ends in the west at the Straits of Gibraltar and does not continue beneath the waters of the Atlantic Ocean.

The geosyncline, during the period of its active existence, was divided into intrageosynclines and intrageanticlines. This segregation cannot be found everywhere, however, since many localities are covered either by the sea or by younger sediments. But from the parts that are visible it may be concluded that the intrageosynclines are closed or semien-closed elongated ovals. Their number in the cross section of the geosyncline and their lateral distribution change in such a manner that it is impossible to extrapolate solid intrageosynclinal and intrageanticlinal zones throughout the full extent of the geosyncline.

An examination of other examples would lead to similar results. One can merely say that the external outlines of the older geosynclines--the Hercynian and Caledonian--are evidently simpler and that they are not so clearly divided into large separate ovals.

Since the intrageosyncline is contained within an enclosed space, the thicknesses of its deposits reach a maximum in the center and thence decrease in all directions. The gradient of the change in thickness, however, is different in different directions: it is low in the longitudinal direction and considerably higher across the geosyncline. Even in the same cross section the gradient may be different at the two edges of the intrageosyncline: asymmetrical intrageosynclines are often encountered, in which the axis of maximum subsidence is displaced toward one edge. Examples will be given below.

Individual intrageosynclines vary in width from several score to 200 or 300 km; a length of about 100 km is evidently the most common. Intrageanticlines are sometimes very narrow, no more than a few kilometers in width. Some authors call them *cordilleras*.

The lengths of intrageosynclines and intrageanticlines usually reach several hundreds of kilometers. The overall width of the whole geosynclinal zone is extremely varied: in the case of the Alpine geosyncline it ranges from 100 to 1,500 km. The older the cycle, the broader the geosyncline.

Platforms have irregularly rounded, sometimes angular outlines (see the map at the end of the book). Subgeosynclines and subgeanticlines are round or oval in shape; they are much less elongated than intrageosynclines and intrageanticlines. In the majority of cases the sub-

geosynclines have oval shapes, and the subgeanticlines, as it were, fill up the gaps between them.

THE COMBINATION OF WAVELIKE UPLIFTS AND SUBSIDENCE

There is always an opposition of regions of subsidence and regions of uplift of the earth's crust during the course of wavelike oscillatory movements on the earth's surface. While the sediments are being accumulated in one place, there is simultaneous erosion in another.

While there were subsidence and deposition within the Russian plain during the Middle and Upper Paleozoic and during the Jurassic, Cretaceous, and Paleogene, Scandinavia was undergoing erosion. During the time of uplift of the Russian plain, in the Middle and Upper Triassic, sediments of that age were being deposited in the Caucasus. In other words, the stratigraphic column contains no gaps that are ubiquitous: the gaps are always local and their area is limited, even though it may be very large.

The sizes of the areas of uplift and subsidence into which the earth's crust is always divided may differ, from almost an entire continent (as with the uplift of the Quaternary period) to a few square kilometers. The shapes of the individual areas are mainly round or elliptical. Even if the area of subsidence or uplift at first glance appears to have a linear, elongated shape, a closer examination will show that it is divided into separate adjacent ovals. Examples will be given below.

The areas of uplift and subsidence are usually linked by a smooth bending of the earth's crust, without breaking its continuity. But where neighboring zones of uplift and subsidence are very close to each other and the movements in them have a great amount of contrast (as may be seen chiefly in geosynclines), they are often separated by vertical faults with a relative upward or downward slippage of the adjacent zones. The movements along these faults have all the properties of oscillatory movements: they develop slowly and gradually in an undulatory manner; that is, the relative displacements change direction with the course of time, and the portions of the earth's crust that are separated by the fault are displaced vertically in opposite directions relative to each other. The result of such oscillatory movements accompanied by rupture is a very abrupt change in the thicknesses of the deposits between zones of uplift and of subsidence, the change moving by steps along the line of the fault. The facies of the sediments also change very sharply along the same line.

Examples of faulted oscillatory movements were pointed out by Robinson (1937)* in the northwestern Caucasus. A series of longitudinal faults was formed in this region as early as the Paleozoic, dividing the earth's crust into a number of narrow plates, which moved slowly up-

ward and downward with respect to each other. Consequently, the sections of the Carboniferous, Permian, and Triassic differ in different zones both in the thicknesses of the individual stages and in the facies of the deposits. I. G. Kuznetsov (1933, 1951) has pointed out similar examples in other parts of the northern Caucasus.

D. S. Kizeval'ter (1948) has made a specially thorough study of the occurrence and the history of faulted oscillatory movements in the northern Caucasus, in the area between the Malka and the Chegem Rivers. He concluded that this region is broken up into very narrow longitudinal zones (in the direction of the parallels of latitude), each of which is several kilometers wide and has its own distinct section and

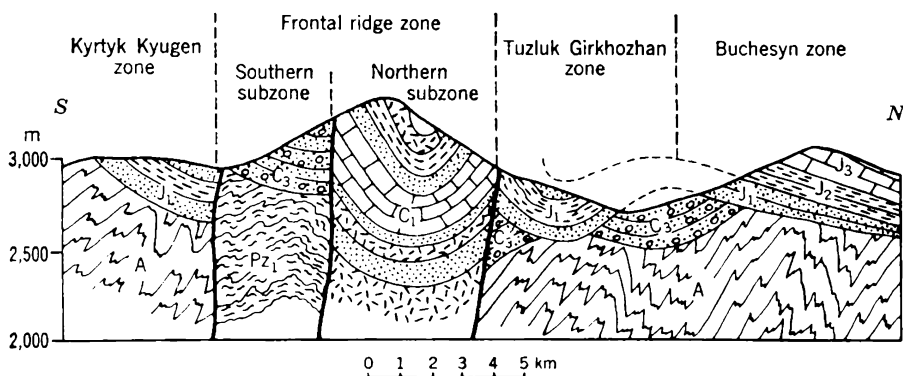


FIG. 114. Diagrammatic geologic section through the frontal ridge of the northern Caucasus: J_3 , Upper Jurassic limestones; J_2 , Middle Jurassic shales; J_1 , Lower Jurassic sandstones; C_2 , Upper Carboniferous sandstones and conglomerates; C_1 , Lower Carboniferous volcanogenic and sedimentary series; Pz_1 , Lower Paleozoic metamorphosed shales; A, Precambrian schists (after D. S. Kizeval'ter).

history. A total of five has been distinguished from north to south: (1) the Buchesyn zone, (2) the Tuzluk-Girkhozhan zone, (3) the northern subzone of the frontal ridge, (4) the southern subzone of the frontal ridge, and (5) the Kyrtyk-Kyugen zone (Fig. 114).

The section of the Buchesyn zone begins with a Precambrian folded base, which is overlain by Mesozoic rocks beginning with the Upper Lias; this is 380 m thick.

The section of the Tuzluk-Girkhozhan zone is made up of the same Precambrian, Upper Carboniferous, and Upper Liassic deposits. The thickness of the Upper Carboniferous is 500 m and that of the Upper Liassic is up to 1,300 m.

The section of the northern subzone of the frontal ridge is composed of Lower Carboniferous and Middle Liassic sediments. The rest of the deposits are absent. The Lower Carboniferous apparently lies directly

upon the Precambrian. The Lower Carboniferous is up to 2.5 km and the Middle Liassic 600 m thick.

Metamorphic rocks of the Lower and Middle Paleozoic lie at the base of the section through the southern subzone of the frontal ridge. These are unconformably overlain by Upper Carboniferous and Permian rocks. The section also contains Upper Jurassic and Upper Pliocene deposits.

In the Kyrtyk-Kyugen zone, Upper Liassic deposits lie directly upon Precambrian schists. Above the former, the section contains deposits of the Middle Jurassic (?) and Upper Jurassic as well as Pliocene lavas.

All these zones are separated from each other by vertical faults. The consolidated section in Fig. 114 shows that along these faults there were long-lasting vertical movements that changed in direction.

After analyzing the sections of the various zones with the purpose of determining the distribution of thicknesses and the facies in the deposits of different ages, Kizeval'ter came to the following conclusions (1948) regarding the general structure and the history of the area:

The two northern and the extreme southernmost zones are distinguished by an almost total absence of Paleozoic sediments and by relatively great thicknesses of the Upper Liassic-Dogger deposits. . . . In complete contrast to these zones there is the frontal ridge zone with its great development of Paleozoic deposits . . . and almost total absence of the Mesozoic, resulting in a very abbreviated section. . . .

The history of the frontal ridge, so far as it is known, may be outlined as follows.

The frontal ridge zone first emerged clearly in the Lower Paleozoic as a zone of intense subsidence. . . . In the middle Paleozoic, subsidence developed mainly in the western part of the area and was displaced northward from its Lower Paleozoic position. Along the southern border a fracture formed, with uplift of the southern flank. . . . In the Lower Coal Measures epoch the axis of the subsidence was displaced still more to the north, and an exceedingly thick stratum of sedimentary and volcanic rock was deposited within a very narrow zone. . . . A certain part of the trough, apparently the axis, began to subside especially rapidly along fractures; uplifts north and, especially, south of the subsiding zone at times produced coarse clastic material. . . .

In the Tuzluk-Girkhozhan zone during the Stephanian age, on the site of some former uplifts which provided the source of the material in the Lower Coal Measures basin, a narrow zone began to subside. . . . To the south, in the area of the recent Lower Coal Measures subsidence, an uplift developed. . . . South of the growing uplift, toward the end of the Upper Carboniferous, a second subsidence appeared in the southern subzone of the frontal ridge and continued into Lower Permian times as well.

This analysis shows that the movements of the individual narrow zones were typically oscillatory and simultaneous with the deposition of sediments, thus influencing the thickness and facies of the deposits. But

these movements took place along the faults. A comparison of the thickness of the earth's crust (which is several score kilometers) with the width of the individual zones (several kilometers) leads one to conclude that the extremely narrow vertical plates of the crust are independent in their movements.

A great amount of attention was devoted to the faults that accompany oscillatory movements by A. V. Peyve (1947*a* and *b*), who described faults of this nature in Tien Shan and the Urals. For these faults he introduced the special term *deep-seated ruptures*; this term will be used henceforth in this book. In his writings, A. V. Peyve considered that the deep-seated ruptures were independent and that they were in

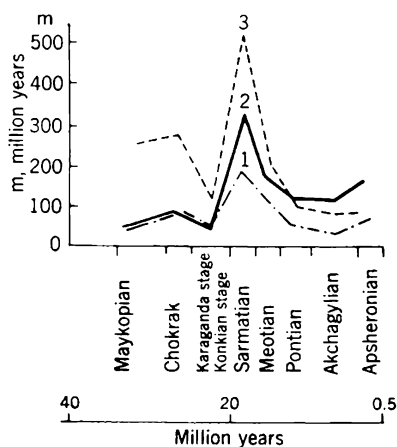


FIG. 115. Changes in the rate of uplift of the Caucasus ridge and of subsidence in the Ciscaucasian and Kura depressions during the Neogene: (1) Tera-Kuban depression (subsidence); (2) Kura depression (subsidence); (3) main Caucasus ridge (uplift) (after A. B. Ronov).

all cases primary dislocations that determine the positions of the zones of uplift and subsidence of the earth's crust. In the present author's opinion, deep-seated ruptures may be classed as both primary and secondary. The primary ruptures will be discussed below. The others are secondary ruptures in the continuity of the earth's crust that result from the great contrast of the oscillatory movements.

The development of subsidences and uplifts located side by side reveals a tendency toward mutual compensation. This is shown in the fact that in certain areas (geosynclines) the zones of intensive subsidence are always accompanied by adjacent zones of uplift with the same degree of intensity, whereas other areas (platforms) are divided into zones of small subsidence and small uplift. Examples of a great subsidence next to a small uplift or of a great uplift next to a small subsidence occur only as rare exceptions to the general rule. In many

cases it may be seen that an increase in the degree of uplift is accompanied by a corresponding increase in the neighboring subsidence, and vice versa; this is particularly clear in the second stage in the development of a geosyncline. At this stage a large uplift appears in the middle of the geosyncline, and at the same time there is a formation on the periphery of frontal flexures, whose subsidence as it were compensates for the central uplift. As an example, Fig. 115 compares the change in the volumetric rate of uplift of the Caucasus ridge during the Neogene with the change during the same time in the rate of subsidence of the Ciscaucasus frontal flexure. It is clear that the changes in these two rates are parallel.

The same conjugation also takes another form: when subsidence in a certain zone changes with the course of time to uplift, a major subsidence, as many examples show, usually changes to a major uplift and a minor subsidence to a minor uplift. In geosynclines one may observe the great amplitude of the subsidence and of the uplift that succeeds it in the same place, whereas on platforms, after a small subsidence, there is a small uplift. These movements in opposite directions that follow after each other are not mirror images, however, and there may be noticeable differences in their magnitudes.

CHAPTER 16

HISTORY OF OSCILLATORY MOVEMENTS

HISTORY AND LAWS GOVERNING THE DEVELOPMENT OF OSCILLATORY MOVEMENTS IN GEOSYNCLINES

The history of oscillatory movements in geosynclines shows that these movements occur with regularity, following a definite succession. Although individual geosynclines differ from one another in their history, common regular features in the development of oscillatory movements can be found in all of them.

The study of oscillatory movements in geosynclines encounters great difficulties in method. The chief impediment results from the fact that complete sections of geosynclines, in which all the interrelationships of facies and thicknesses may be observed, are comparatively rare. Usually only part of the geosyncline can be observed. It is especially difficult to study the central parts of the geosyncline, which are often sunk and are either covered by the sea or buried by later deposits. Deep erosion in mountainous regions destroys whole formations, so that it is difficult to reconstruct their structure and facies.

Because of these difficulties, the mechanism of the oscillatory movements in geosynclines can be determined only by comparing the results of many separate observations made in different geosynclines, in different parts of geosynclines, and for different cycles. Hence everything is still far from clear in the development of oscillatory movements in geosynclines, and various researchers understand the problems of oscillatory movements differently. Nevertheless, recently, owing to the increased work of researchers, many obscurities have been cleared up and differences of opinion eliminated.

At first the mechanism of oscillatory movements in geosynclines was established from the example of the Caucasus; thus it is convenient to begin this survey with the Caucasus geosyncline. Before attempting a description of the concrete examples, the general course of development of oscillatory movements in geosynclines will be presented. Although the opposite order of study would have been more logical, it is difficult for this book to reflect the preliminary scientific analysis of the factual

material. The reader can only be convinced of the correctness of this scheme of geosynclinal development after having worked independently with the data; the present author's task is to show the path he believes must be followed in order to solve this problem.

General Scheme of Development of Oscillatory Movements in Geosynclines

Geosynclinal zones have existed on the earth during all the geotectonic cycles known to us. Development of the geosyncline at present has two aspects, which must be kept equally in mind. In the first place, it is a directional development occurring during the entire geologic history of all the geotectonic cycles; it reflects a general progression in the development of the structure of the earth's crust. In the second place, it is a cyclic repetition of similar events corresponding to the general periodicity in the oscillatory movements of the earth's crust. This periodicity is closely related to geotectonic cycles and is one of the ways in which they are manifested.

The superimposition of a periodicity on a progressive, unidirectional development gives the entire geotectonic process a cyclic nature. There is a periodic return to similar situations at equivalent stages of different cycles, while the changes continue uninterruptedly to increase from one cycle to the next.

This section will describe the aspect of the geosyncline's "life" that is determined by the periodicity of the geotectonic process. On the basis of the similarity between the different cycles, the attempt will be made to establish the particular mechanisms in the development of the geosyncline during each cycle. Such mechanisms must inevitably be comparative and tentative: they must change with the course of time under the influence of the directional changes taking place in the tectonic processes. But observation confirms the supposition that, for the last three major cycles, these changes do not affect the accuracy of the determinations of structures that can be made at present. Thus it can be said that during the Caledonian, Hercynian, and Alpine cycles approximately the same succession of movements of the earth's crust took place in the geosynclines.

Because of this periodicity, the Caledonian, Hercynian, and Alpine geosynclines can be treated separately; on the other hand, it must at the same time be kept in mind that this division reflects only one aspect of the phenomenon. The other aspect is due to the fact that the geosynclines of different cycles are only stages in a single development of geosynclinal conditions on the earth's surface.

In brief, the mechanism of geosynclinal development during the cycle may be formulated as below (Belousov, 1948; W. Belousoff, 1956).

Throughout the whole cycle, the geosyncline is divided into areas of intensive subsidence and uplift of the intrageosynclines and intrageanticlines adjoining one another, with great contrast in the amplitude of their movements in opposite directions. These particular zones, with elongated, oval outlines, extend parallel to each other along the trend of the geosyncline as a whole or else are disposed diagonally in relation to the latter and en echelon in relation to each other.

As regards the general periodicity of the oscillations, because of the superimposition of the oscillatory movements on the wavelike process, the history of the geosyncline during a cycle is divided into two stages. Subsidence predominates during the first stage; it is manifested in the slow and irregular expansion of the intrageosynclines at the expense of the intrageanticlines separating them, whereas on the periphery they expand at the expense of the adjacent platform. Some intrageanticlines are also involved in the subsidence, and sediments are deposited on them; but this subsidence is very much less intensive than that of the intrageosynclines. The intrageanticlines are preserved as zones of relative uplift. During the second stage, again in relation to the general periodicity of the movement, the predominance passes to uplift. Now the surface area of the zones of subsidence slowly decreases, while the zones of uplift slowly increase and the uplift becomes more and more extensive. It will be noted, in anticipation, that the uplift is accompanied by intensive folding movements. Finally, in place of the geosyncline, high mountain ranges appear that are separated by intermontane basins, and the geosyncline thus becomes a complexly dissected mountain region.

In some cases, at the very beginning of the cycle, one may distinguish an *initial substage*, or *epoch*, in the *formation* of the geosyncline, characterized by coarser division of the geosyncline into areas of subsidence and uplift. In such cases, the impression is created that at first the geosyncline forms as the result of a number of very large, individual depressions, which later, but still at the beginning of the first stage, separate into narrower intrageosynclines after the uplift of new intrageanticlines within these depressions. Thus during the first stage, complications appear in the structure of the geosyncline, and its internal division becomes more detailed.

In every case, at the very end of the cycle one can distinguish a *substage*, or *epoch*, of *mountain building* corresponding to the formation in geosynclines of high mountain ranges separated by deep, intermontane basins not covered by the sea. Here "mountain building" is used in its straightforward morphologic meaning, which is the only correct one. A strong protest must be made against the use of this term in the Soviet specialized literature as a translation of the unsuitable and obsolete foreign term *orogenesis*, combining the concepts of folded

and faulted dislocations. Here the present author has in mind only the *formation of mountains* resulting from anticlinal uplift of the earth's crust during the course of its oscillatory, wavelike movements.

It will be seen from the preceding discussion that the second stage of the geosynclinal cycle is separated from the first by a change in the tectonic conditions or a reversal of them—a transition from predominant subsidence to predominant uplift. This is called the *general reversal*, or *general inversion*, of geotectonic conditions. The transition from the first stage to the second, however, is not marked by that feature alone; it is also characterized by a redistribution of the zones of subsidence and uplift. The centers, or rather the axes, of this redistribution are the intrageosynclines. New, secondary uplifts, which will be called *central uplifts*, are formed in them during the transition from the first stage to the second, at different times and in different places. Each of these central uplifts divides the intrageosyncline into two smaller depressions (*marginal basins*). Later one may observe a gradual growth and expansion of the central uplift coinciding in time with an outward displacement of the marginal basins, which in moving apart “roll,” so to speak, upon the neighboring intrageanticlines, gradually drawing them by their edges into the subsidence.

If this process goes far enough and if the intrageanticline is relatively narrow, then toward the end of the cycle a complete reversal of the disposition of the zones of uplift and subsidence occurs in the geosyncline. The central uplifts, by expanding, completely replace the former intrageosynclines, as if forcing them out; and in the place of the former intrageanticlines, new depressions result from the outward displacement of the marginal basins: the two marginal basins moving the single intrageanticline from both sides meet, merge, and become a single intermontane basin. If, however, the intrageanticline is sufficiently wide, it incorporates the marginal basins at their edges while its central part remains a zone of steady elevation until the end of the cycle. On the periphery of the geosyncline the marginal basin “rolls” upon the edge of the platform and is transformed into a so-called fore-deep. The transformation of a particular subsidence (intraegeosyncline) into uplift is called *particular reversal*, or *particular inversion*, of geotectonic conditions.

Thus the geosyncline presents a picture of wavelike development of oscillatory movements: within the wave of subsidence (intraegeosyncline) there appears a wave of uplift (central uplift), after which the two waves expand—both the subsidence and its superimposed central uplift. This process occurs under the conditions of a combination of uplift and subsidence: the subsidence of the marginal basins that border the central uplift on both sides “compensates” the latter, and its rate and extent depend directly on this uplift.

The active development of the geosyncline in the second stage is focused on the intrageosynclines, where new uplifts are formed. The intrageanticlines appear to be the more stable elements of the geosyncline; they tend to be preserved and to change into zones of subsidence only when the marginal basins of the neighboring intrageosynclines "roll" upon their edges.

The difference in development between individual intrageosynclines within the same geosyncline lies in the ability of the wavelike process to run its course to different degrees of completeness and thus to produce different results at the end of the cycle. In one case, when the development runs to completion, a particular inversion is observed and the intrageosyncline is fully replaced by an uplift. In another case, the development is less complete, and although the intrageosyncline is complicated by fairly large central uplifts within it, it is not completely replaced by them. Finally, in a third case, the intrageosynclines are so little complicated by central uplifts that they are maintained in their essential features (like the intrageanticlines in this case) throughout the entire cycle.

An attempt will be made later to show that all these different cases are not disconnected but are regularly connected with each other. The primary case is that of the full development through a particular inversion. Cases of incomplete development are reflections of a regular transition from geosynclinal to platform conditions: throughout the history of the earth's structure, the platforms succeed geosynclines. Regions in which the development is incomplete but which are nevertheless not typical platforms will here be called *parageosynclines*. In Soviet literature the term *sempiplatform* has also been used for such areas (Sapozhnikov, 1948).

Other differences in the development of geosynclines and the zones within them take the form of nonsimultaneous appearance of central uplifts in the intrageosynclines and of asymmetry in the development, when, for example, one of the two marginal basins formed in one intrageosyncline migrates toward the side faster than the other or when the central uplift is not formed in the axial part of the intrageosyncline but is pushed against one of the adjacent intrageanticlines as if piling it up at the side of the intrageosyncline. Other particular features of the development of oscillatory movements will become clear below in the discussion of specific examples.

This scheme of development of oscillatory movements in geosynclines still lacks some important elements. In the substage of mountain building one can in many cases observe the formation of new depressions in the geosyncline; these are of a special nature and are not tied to the previous depressions by the development of wavelike oscillatory movements. These depressions occur in two forms. The first type has a relatively

small surface often bounded by faults formed on the central uplifts and complicating their structure. In the second case, these depressions have very large surfaces and irregular outlines in plan. They are superimposed on the geosyncline, with its interior zones of uplift and subsidence, and absorb considerable parts of them.

The depressions of the first kind, which are of small size and formed on the central uplifts, may be called *superimposed troughs*; the term was suggested by N. S. Shatskiy (1938). For the larger depressions (of the second type), the term *interior depressions* may be used.

Superimposed troughs with round or oval outlines (often broken, because they are bounded by faults) form in various places in the trend of the central uplift and are not related to each other. Appearing in the axial zone of the central uplift, the superimposed trough separates into two "residual uplifts." But these troughs may form on the flanks of the central uplift and occupy an off-center position.

Within large interior depressions there are unaffected areas of the rising geosyncline, which will be called *marginal uplifts*. These do not belong to the category of tectonic zones, which includes the central uplifts, intermontane basins, and foredeeps. Because the interior depressions are placed discordantly on the geosyncline, independently of its internal structure, the marginal uplifts include heterogeneous elements of the geosyncline and may consist of parts of central uplifts and of stable geanticlines irregularly cut by the boundaries of the interior depression. In some regions the interior depression may overlap all the marginal zones of the geosyncline and immediately touch the adjacent platform. Thus "marginal uplift" is a term introduced only to designate the part of the geosyncline that has remained outside the limits of the interior depression.

The subsidence of the earth's crust in superimposed troughs and interior depressions occurred during the preceding stages of the geosynclinal development. In the superimposed troughs and interior depressions sedimentary strata are deposited unconformably on the rocks that form the central uplifts and intermontane basins of the geosyncline. These sedimentary rocks are relatively thin and, what is most important, their gradients are small, since moderate thicknesses are combined with relatively large areas of the basins. These tectonic structures thus approach the depressions of the platform type—synclises.

This conclusion is supported by the fact that the deposits of superimposed troughs and interior depressions are not intensively folded; for the most part they form horizontal strata broken only by discontinuous, domical folds and bending of the beds at the edges of the depressions associated with displacement along normal faults. For this reason, superimposed troughs and interior depressions must evidently be regarded as tectonic structures that are not so much geosynclinal as

transitional between geosynclinal conditions and platform conditions (see below). This formation within the geosynclines of synclises properly belonging to platforms reflects the "disintegration" of the geosynclines. But superimposed troughs and interior depressions are formed during the stage of mountain building, while outside them, in the parts of the geosyncline that remain, movement continues on a large scale. It will be seen below that the formation of an interior depression does not prevent the continuation of folding beyond its limits.

Thus although the superimposed troughs and interior depressions already show a partial transition from geosyncline to platform in space, in time this process begins in the period of mountain building. Platform and geosynclinal conditions merge into each other, so to speak, both in space and in time. Thus it has been considered expedient to consider these tectonic formations in the general scheme of geosynclinal development.

Examples of superimposed troughs are certain small depressions formed in the Devonian system on the eastern slopes of the Urals upon the Caledonian central uplift; the Upper Triassic and Lower Jurassic brown-coal basins in the Chelyabinsk region that were formed during the Hercynian mountain-building epoch in the Urals; similar small lignite basins of Late Carboniferous age formed in the interior zone of the Hercynian geosyncline in Western Europe, also at the time of Hercynian mountain building; a number of depressions in central Kazakhstan; and others. A typical example of a large interior depression is the western Siberian lowland, a vast and complex syncline formed during the beginning of the Jurassic on the site of the extensive Ural-Siberian geosyncline. This depression covered a large number of interior zones of the geosyncline: central uplifts, intermontane basins, and stable intrageanticlines. To the west, outside the interior depression, there remained the Urals range, consisting of a number of heterogeneous tectonic zones and appearing to be a marginal uplift. To the east, the interior depression immediately adjoins the Siberian platform for a considerable distance. Separate fragments of the marginal uplift are seen in the Salair and the Tom'-Kolyvan arch and to the north in the mountains of the Taymyr peninsula. A number of other examples of interior depressions will be given later.

This scheme of the development of a geosyncline is represented graphically in Fig. 116, which should be examined from bottom to top. At the beginning of the cycle (diagram I) the geosyncline represented here was divided into four areas of subsidence and three uplifts between them, and the thicknesses of the accumulated strata were distributed accordingly. The areas of subsidence expanded gradually at the expense of the uplifts and of the adjoining platforms (diagram II). Later central uplifts were formed in the two intrageosynclines on the right; during

this process, in the intrageosyncline farthest to the right, the central uplift developed asymmetrically and combined with the adjacent intra-geanticline, whereas in the next intrageosyncline the central uplift occupied the axial position (diagram III). Still later a central uplift was formed in the depression second from the left as well, and all the central uplifts underwent considerable growth. At the same time the marginal basins "rolled upon" the neighboring intrageanticlines. The

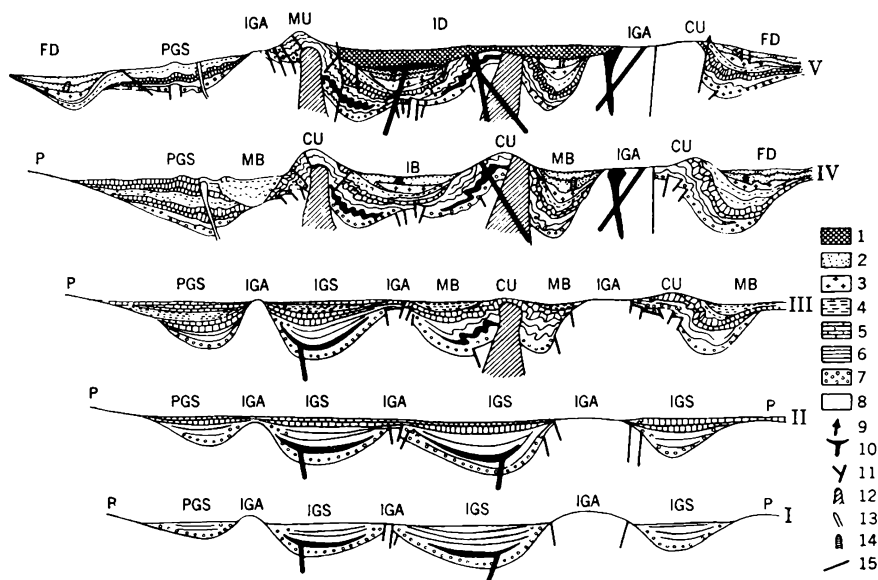


FIG. 116. Diagram of the development of a geosyncline during a cycle: I-V, stages of development of the geosyncline; P, platform; PGS, parageosyncline; IGS, intra-geosyncline; IGA, intrageanticline; CU, central uplift; IB, intermontane basin; FD, foredeep; MB, marginal basin; ID, interior depression; MU, marginal uplift; (1) interior depression formation; (2) molasse formation; (3) lagoonal formation (salt, gypsum); (4) upper terrigenous formation; (5) limestone formation; (6) lower terrigenous formation; (7) coarse clastic deposits; (8) rocks of preceding cycles; (9) volcanos; (10) extrusives and sills; (11) fissure intrusives; (12) batholiths; (13) minor intrusions; (14) diapirs; (15) tectonic faults.

central intrageanticline, because it was narrow, was entirely involved in the subsidence, and an intermontane basin was formed on its site, while the neighboring intrageosynclines were totally replaced by uplifts (diagram IV). Here the conditions were completely reversed: since the right intrageanticline was broader, its edge was drawn into the subsidence. Finally, the basin farthest to the left did not undergo partial inversion; here parageosynclinal conditions obtained. The uppermost diagram (V) depicts the epoch of mountain building, when superimposed troughs and interior depressions were formed along with the uplift of the mountains.

This diagram, besides the oscillatory movements, also depicts other types of movement—folding and faulting—as well as igneous activity.

The scheme presented above is one reflecting the average line of development. It shows that the different variations considered correspond to particular deviations from this average line and cause the complications in it. The basic process of wavelike oscillatory movements is complicated by oscillations of secondary and lesser order, which reinforce each other but do not distort the process to the degree that the basic regularities in it can no longer be traced with sufficient clarity.

The following circumstances are extremely important. In this book the oscillatory movements were divided into wavelike and general oscillations. The first are reflected in the development of uplift and subsidence—of intrageosynclines and intrageanticlines. The second determine the periodicity in the development of the earth. It can be seen from the scheme presented here that these two classes of oscillatory movements do not run their course independently of each other but are closely related to such an extent that the specific stages of the cycle are characterized by definite stages in the development of uplifts and areas of subsidence. Some observations on this subject will be made in the final chapter. Below, since they have been most thoroughly studied, the development of oscillatory movements in the Taurus-Caucasus geosyncline will be examined in greater detail, and some information will be given briefly on the development of certain other geosynclines.

Oscillatory Movements in the Taurus-Caucasus Geosyncline of the Alpine Type. Between the Kuban steppes to the north and the Arabian deserts to the south lies a region known to have been extremely mobile throughout the whole of geologic history: the Taurus-Caucasus geosyncline. The most thoroughly studied is the last, or Alpine, cycle of its existence, including the Mesozoic and Cenozoic. Here the development of the Taurus-Caucasus geosyncline in the Alpine cycle will be discussed.

A large number of works by Soviet researchers are devoted to the geology and tectonic history of the Caucasus. The history of the oscillatory movements of the earth's crust in the Caucasus, on the basis of analyses of the available data that throw some light on the distribution of the facies and thickness of the sedimentary strata, has been studied especially by V. V. Belousov (1938–1940), M. I. Varentsov (1950), M. V. Gzovskiy (1948), L. N. Leont'yev (1950), M. V. Muratov (1949), V. Ye. Khain (1950), and others. The discussion here will be based on these works. The parts of the Taurus-Caucasus geosyncline outside the Soviet Union have been little studied. Nevertheless, the summary work by Arni has become available (1939) and will be utilized.

The Alpine cycle in the development of the Taurus-Caucasus geosyncline begins in the Early Jurassic. In this geosyncline the Lower Jurassic

deposits everywhere lie upon more ancient rocks with a sharp unconformity that shows traces of deep erosion. Apparently, before Jurassic time this entire region underwent an uplift that concluded the preceding Hercynian cycle. New subsidence of the earth's crust and marine transgressions associated with it began in the Early Jurassic.

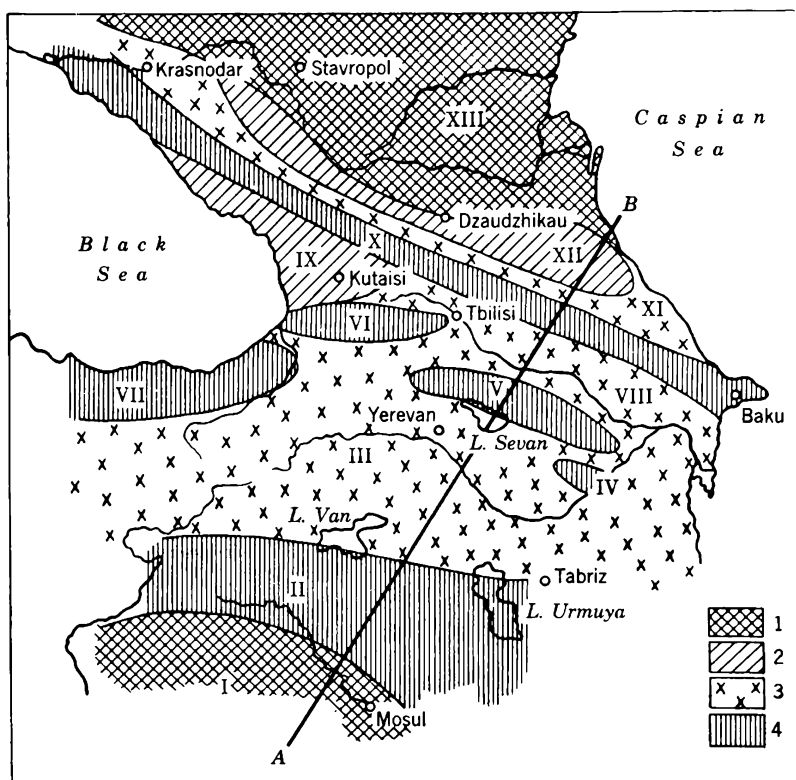


FIG. 117. Diagram of the structure of the Taurus-Caucasus geosyncline before the general inversion: (1) platforms; (2) parageosynclines; (3) intrageanticlines; (4) intrageosynclines; (I) Arabian P; (II) Taurus IGS; (III) Araks IGA; (IV) inferred Kafansk IGS; (V) Somkhet IGS; (VI) Trialet IGS; (VII) Anatolian IGS; (VIII) Kura IGA; (IX) Rion PGS; (X) main Caucasus IGS; (XI) Sadonian IGA; (XII) northern Caucasus PGS; (XIII) Russian P; A-B, line of section (see Fig. 118).

The distribution of facies and thicknesses of the Mesozoic era in the Taurus-Caucasus geosyncline was divided into several intrageosynclines and intrageanticlines. The limits of the geosyncline were the Arabian platform to the south and the Russian platform to the north. These platforms underwent slight subsidence during the Mesozoic, but in the parts of these platforms next to the geosyncline the predominant movement was uplift. These were the sources of the clastic material deposited in the neighboring regions.

In moving northward from Arabia, within the boundaries of the Taurus mountain region, one observes an increase in the thickness of the Mesozoic deposits. On the Arabian platform, the Jurassic and Cretaceous deposits are very thin or entirely absent, whereas in the Taurus region the total thickness of the Mesozoic rocks is more than 8 km. Thus the younger strata of the Mesozoic, the Upper Cretaceous, spread farther over the platforms than the older strata. Northward from the Taurus Range toward Lake Van, the Mesozoic thicknesses decrease

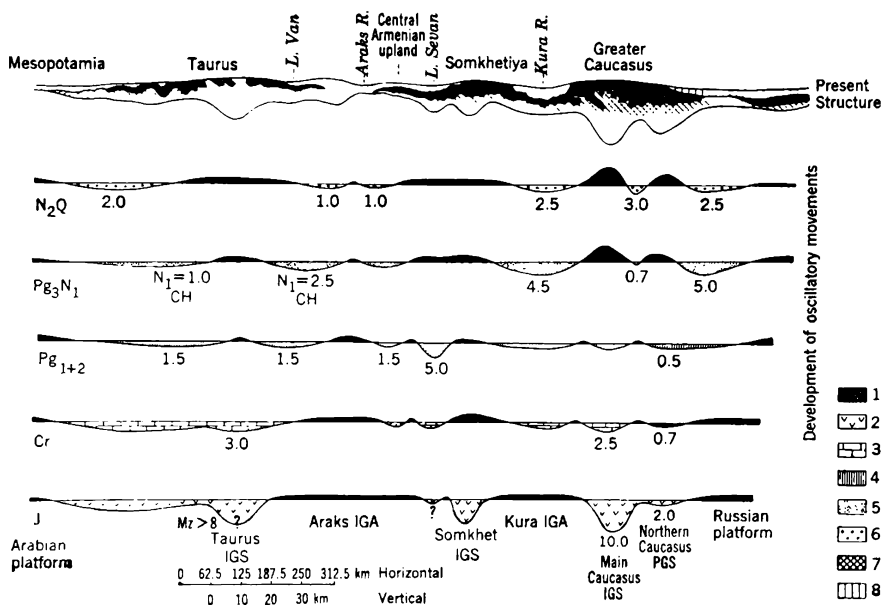


FIG. 118. Diagram of the development of the Taurus-Caucasus geosyncline: (1) on all graphs except the top, uplift; on the top graph, Mesozoic; (2) Jurassic; (3) Cretaceous; (4) Eocene; (5) Oligocene, Miocene, and Quaternary rocks; (7) pre-Mesozoic rocks; (8) Tertiary rocks. Numbers are approximate thicknesses in kilometers. Lower line on top graph is maximum subsidence of the base of the Mesozoic before inversion.

again, while still farther north the Lake Van Triassic, Jurassic, and Lower Cretaceous deposits wedge out completely and the section contains only thin, unconnected blocks of Upper Cretaceous deposits lying directly on the Paleozoic.

Thus a traverse from the southern edge of the geosyncline to Lake Van will cross the Mesozoic intrageosyncline, which is called the *Taurus intrageosyncline* (Figs. 117 and 118), and enter the territory of the intrageanticline called the *Araks intrageanticline*. This intrageanticline is extremely broad: its southern edge is at Lake Van; to the north it includes all the southern and central part of the Armenian S.S.R. (the

Araks Valley and the central Armenian upland), where Mesozoic deposits are represented by thin horizontally disconnected beds of mainly Upper Cretaceous rocks lying on Paleozoic and Precambrian rocks. This region is distinguished by its extremely intensive Late Tertiary and Quaternary volcanic activity, and the young lavas to a great degree conceal the structure of the intrageanticline.

Farther north the situation changes again. In the region of Lake Sevan and in the Somkhet-Karabakh Mountains on the northern edges of the Lesser Caucasus, thick deposits of Jurassic and Cretaceous rocks appear. The Lower and Middle Jurassic are represented here by sandy-argillaceous strata and by a series of tuffs and lavas, whereas the Upper Jurassic consists mainly of limestones. The Lower Cretaceous is nearly absent, but the Upper Cretaceous is represented by both limestones and thick volcanic strata of tuffs and lavas. The Jurassic and Cretaceous deposits are somewhat separated in space: the Jurassic deposits are developed along the axis of the Somkhet-Karabakh Mountains, whereas the Upper Cretaceous is located mainly farther to the south of them, near Sevan and on the northern slopes, which incline downward toward the Kura depression. The Jurassic deposits are several kilometers and the Cretaceous deposits two or three kilometers thick. This area is thus within the limits of another intrageosyncline. The observed decrease in the thickness of the Jurassic deposits in the Somkhet-Karabakh Mountains, moving toward the Kura depression, shows that during the Jurassic period the intrageosyncline apparently occupied only a narrow strip of the mountains, whereas in the Cretaceous period it widened considerably southward toward Lake Sevan and northward within the limits of the Kura depression. This is called the *Somkhet intrageosyncline*.

The diagram in Fig. 117 shows a separate Kafan intrageosyncline southeast of Lake Sevan. Here there are thick deposits of Jurassic age. It is possible, however, that this intrageosyncline comprises a single unit with the Somkhet intrageosyncline.

North of the Somkhet intrageosyncline, on the site of the present Kura depression, there was an intrageanticline (the *Kura intrageanticline*) in Mesozoic times. The existence of this intrageanticline has been established from the change in the facies and thicknesses of the Mesozoic deposits on the southern slopes of the main Caucasus range. Here in many sections one can trace the rapid decrease in the thickness of the Jurassic and Cretaceous deposits during its lateral displacement toward the Kura depression, and the change toward an increase in the amount of clastic and coarse clastic material. This indicates the presence of uplifts and areas of erosion within the territory of the present Kura lowland. The eroded areas were probably greater during the Jurassic period and decreased considerably during the Late Cretaceous, since

the thicknesses of the Upper Cretaceous are considerable on the northern slopes of the Somkhet Mountains.

Upper Cretaceous sediments were apparently deposited on the Kura intrageanticline, but not everywhere. In the foothills of the southern slopes of the main Caucasus range, for example, there was a narrow zone of erosion during the Late Cretaceous. To the west the Kura intra-geanticline bordered directly two zones: the *Trialet intrageosyncline* and the *Rion parageosyncline*. In the Trialet zone Jurassic rocks are nowhere exposed in the area of the present Adzharo-Trialet range; whether they were deposited here remains unknown. In any case there was deposition during the Late Mesozoic, when carbonate and volcanogenic strata of great thickness were formed here. The continuation of the Trialet intrageosyncline in Turkey is the *Anatolian intrageosyncline*, which will not be examined here.

The Rion zone during the whole of the Mesozoic was an area of considerable subsidence of the earth's crust. The total thickness of the Mesozoic deposits in the central part of the zone, where the greatest subsidence took place, apparently reaches several kilometers. A great variety of rocks—from clastic rocks to limestones and volcanic rocks — were deposited in this depression. The parageosynclinal nature of this depression will be established below.¹

Further north we enter the area of the main Caucasus range (the Greater Caucasus). Examination of the facies and thicknesses of the Lower Jurassic here shows that at that time the Greater Caucasus was a single large basin in which were deposited very thick sediments (up to 5 or 6 km). This was bounded on the north by the uplift of the Russian platform and on the south by the Kura intrageanticline and by a narrow intrageanticline extending between the basin of the main Caucasus range and the Rion parageosyncline. But as early as the Middle Jurassic the separation became more complicated: the single depression is divided into two zones containing very thick accumulations of Mesozoic deposits.

One zone extends chiefly along the southern slopes of the main Caucasus range but also includes the watershed, as can be seen very well in the southeast and in the northwest at both plunging ends of the range. In this zone the thickness of the Mesozoic deposits is very great; the total thickness exceeds 10 km. The facies of these Mesozoic deposits are extremely varied: they include the sandy-argillaceous beds of the Lower and Middle Jurassic, the carbonate and terrigenous flysch of the Upper Jurassic and Cretaceous, and so forth.

The second zone of great thicknesses of Mesozoic deposits extends mainly along the lower part of the northern slopes of the main Caucasus

¹ According to the most recent data, geanticlinal conditions similar to those of the Kura depression existed in the Kolkhida region.

range. Here the Mesozoic deposits are thinner than in the southern zone, but they still reach 3 to 4 km in thickness. The rocks are varied: sandy-argillaceous deposits of the Lower and Middle Jurassic; limestones, red-colored rocks, and gypsum of the Upper Jurassic; terrigenous deposits of Lower Mesozoic; and limestones and marls of Upper Mesozoic.

Thus we have two intrageosynclines. Between them runs a narrow belt in which the Mesozoic rocks are of small thickness. This zone is seen clearly at the ends of the range, in southern Dagestan—in the Shakh-Dag region and in the Kusar inclined plain—and in the area of the Paleozoic front range in the northwestern Caucasus. In the central part of the range this zone may be seen in the region of the Sadonian deposits. Within this zone the thickness of all the Mesozoic deposits is much smaller; in some places they wedge out almost completely, and in others their thickness decreases to a few hundreds or scores of kilometers. Here there are also coarser clastic facies indicating the presence, within this zone, of areas of absolute uplift and erosion.

Hence, beginning with the Middle Jurassic, within the main Caucasus range one can distinguish, from south to north, the main Caucasus intrageosyncline, the Sadonian intrageanticline, and the northern Caucasus parageosyncline; the parageosynclinal character of this depression will be seen below. The northern Caucasus parageosyncline to the north borders on the Russian platform. In summary, the Taurus-Caucasus geosyncline as a whole in the Mesozoic contained the following zones: (1) the Taurus IGS, (2) the Araks IGA, (3) the Somkhet (Kafan) IGS, (4) the Kura IGA, (5) the Trialet (Anatolian) IGS, (6) the Rion PGS, (7) the main Caucasus IGS, (8) the Sadonian IGA, and (9) the northern Caucasus PGS.

Nearly everywhere the Upper Mesozoic (mainly Upper Cretaceous) deposits are more widespread than the Lower Mesozoic. The Upper Cretaceous in many places extends beyond the limits of the intrageosynclines and overlaps the intrageanticlines. Thus in the Mesozoic development of the geosyncline there was a predominance of subsidence and an expansion of the intrageosynclines at the expense of the intrageanticlines.

The Tertiary deposits are disposed throughout the geosyncline in an entirely different manner. Their thicknesses are greatest where the Mesozoic is thin and nonexistent where the Mesozoic deposits are thickest.

In the Taurus area sandy-argillaceous Eocene deposits of a flyschlike character lie on the slopes of the range. The Oligocene and Upper Tertiary deposits, represented by coarse clastic facies, are of considerable thickness (up to 5 km) in two zones, north and south of the Taurus, at the northern edge of Arabia and in the Araks River basin, where Tertiary rocks locally lie directly on Paleozoic. The distribution of coarse

Tertiary deposits shows that the Taurus range already existed at that time and was the source of the clastic material. A great accumulation of Eocene deposits a few kilometers thick is observed in the region of Lake Sevan. The Eocene rocks, which are also distinguished by their great thickness, are represented by marls, limestones, and sandy-argillaceous suites and by volcanic formations in the Trialet zone. Equally thick Eocene rocks are also developed in the Rion PGS (except Kolkhida).

Oligocene and Upper Tertiary clastic rocks of great thickness were deposited in the Kura and also in the Rion depressions. Here their thickness probably reaches 10 km in places. Another zone of equally great thicknesses of Oligocene and Upper Tertiary clastic deposits extends along the northern foothills of the main Caucasus range. From the distribution of the facies it appears that at that time the main Caucasus range was intensively uplifted and that clastic material was transported from it in great quantities both toward its northern foothills and toward the Kura lowland.

Comparison of the distribution of the Tertiary and Mesozoic deposits shows that between the Mesozoic and Cenozoic eras in the intrageosynclines of the region under consideration there was a partial inversion and that new areas of subsidence appeared alongside the central uplifts. The time of inversion differs in different intrageosynclines; moreover, it cannot be accurately established everywhere. In the Taurus intrageosyncline the inversion apparently took place before the Eocene. Overlying the limestone-marl rocks of the Upper Cretaceous, the sandy-argillaceous Eocene deposits are encountered only on the flanks of the Taurus, which were apparently already being uplifted. Later, in the Oligocene and the Neogene, this uplift on the site of the Taurus intrageosyncline widened, but the areas of subsidence at its foothills were displaced farther to the side. In the south, a typical foredeep formed, and to the north, in the Araks basin, an intermontane basin was filled with Tertiary sediments.

In the Somkhet intrageosyncline the inversion occurred at least before the Late Cretaceous, although in places it may have taken place earlier, even before the Late Jurassic. The Lower and Middle Jurassic deposits are thickest in the axial zone of the Somkhet-Karabakh range. But before the Late Cretaceous this axial zone had undoubtedly undergone uplift, and Upper Cretaceous and Eocene sediments were deposited more to the north and south—in the Kura depression on one side and in the Sevan depression on the other. Here marginal basins were formed and were gradually displaced to the side.

The inversion in the Trialet area probably occurred before the Oligocene. At that time the inversion in the main Caucasus intrageosyncline took place: here a central uplift appeared, bordered on the

north and south by depressions being filled with Tertiary deposits. To the south lies the Kura intermontane basin and to the north the Cis-caucasus foredeep.

The Rion and the northern Caucasus depressions did not undergo the usual inversion. The Rion basin continued to subside until the end of the cycle and in the Late Tertiary became an intermontane basin like the Kura depression, with which it formed a single intermontane basin zone. The northern Caucasus depression was "passively" involved in the uplift during the Late Tertiary, which also saw a gradual widening to the north of the central uplift in the main Caucasus intrageosyncline. The lack of a proper inversion assigns both of these depressions to the category of parageosynclines.

It should be noted that the facies of Tertiary deposits, beginning with the Oligocene and ending with the Pliocene, change throughout this geosyncline from marine and argillaceous to coarser and continental deposits. This testifies to the gradual increase of uplift compared with subsidence. During the Quaternary the region was elevated in the central uplifts, where mountain ridges were formed, but depressions were preserved in the foredeeps and intermontane basins. Thus the general trend of development of the geosyncline in the Tertiary was the opposite of its trend in Mesozoic times: then subsidence was predominant, whereas now it is uplift that predominates. Thus one may speak of the appearance of a general inversion in the middle of the cycle in the Taurus-Caucasus geosyncline.

It is easy to see that the characteristics described earlier were manifested in the development of this geosyncline: the wavelike development of oscillatory movements, general and local inversions, the formation of central uplifts and marginal basins and the outward displacement of the latter, and their transformation into foredeeps and intermontane basins.

The substages of formation of the geosyncline (only in the case of the Greater Caucasus) and of mountain building have been distinguished. In the Early Jurassic there occurred the substage of formation of the geosyncline in the Greater Caucasus when the geotectonic dissection of the region was coarser and when there was only one depression on the site of the main Caucasus range. During the Middle Jurassic, through the appearance of the new Sadonian intrageanticline, this depression was divided into two intrageosynclines. The substage of mountain building occurred at the very end of the Tertiary and in the Quaternary, when the entire geosyncline became a mountainous region.

The individual inversions in various intrageosynclines occurred at different times. Because of the difference in time of the inversions, a certain asymmetry in the development may be observed. When the axial part of the Somkhet zone was being elevated, marginal basins where

Upper Mesozoic and Paleogene deposits accumulated were symmetrically formed in its foothills. During further expansion of the central uplift, the southern (Sevan) marginal basin was displaced southward in the typical manner, and the area of its original location was drawn into the uplift. The northern marginal basin was on the site of the present Kura depression. It did not move farther north but remained in place, apparently because at that time the inversion in the Main Caucasus range was beginning; from there the marginal basin was shifted southward, also descending into the Kura depression.

It can be seen that after inversion, when the central uplifts have grown, the marginal basins moving toward each other within the geosyncline sometimes do not change the intervening intrageanticlines completely into zones of subsidence but leave uplifted areas within their

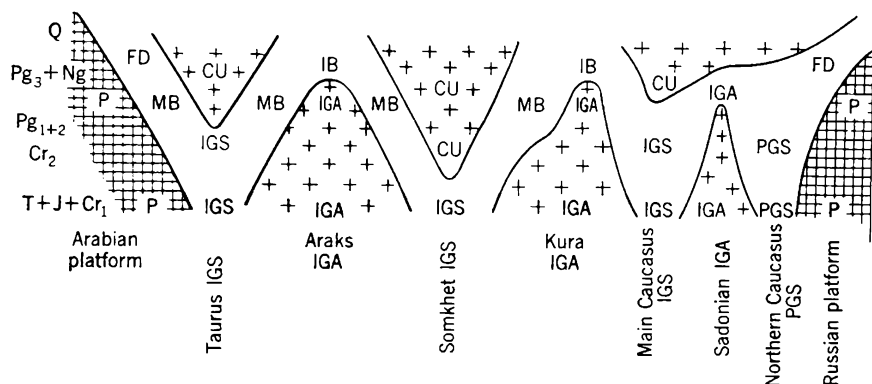


FIG. 119. Development of the Taurus-Caucasus geosyncline, plotted on coordinates representing the cross sections of Fig. 118 and time.

limits, which are preserved during the whole cycle. This happens in the case of broad intrageanticlines. The Araks intrageanticline is an example. It is enclosed by marginal basins on both sides on the Sevan side and on the side of the Araks valley—whereas its central part, corresponding to the central Armenian upland, maintains its regime of uplift to the end of the cycle.

The development of the Taurus-Caucasus geosyncline is represented schematically in a series of stratigraphic sections in Fig. 118. Figure 119 shows the development of this geosyncline in the form of a diagram: if one moves a horizontal line on this diagram from bottom to top, the zones it crosses and their positions will correspond to the regions of uplift and subsidence at the corresponding stage of the geosyncline's development. Thus one can see how, before the inversion, intrageosynclines were expanding at the expense of the intrageanticlines; how, from the time of the inversion, each intrageosyncline "splits" into two marginal basins between which a central uplift appears; how, during the expan-

sion of the latter, the marginal basins move away from each other, narrowing the intrageanticline, and finally merge two by two to form intermontane basins. On the periphery of the geosyncline the foredeeps are formed in the same manner.

This scheme of development of the Taurus-Caucasus geosyncline reflects the basic tendency of the process. In reality, the development is greatly complicated by numerous individual phenomena, although they do not alter the fundamental process. The boundaries of the regions of uplift and subsidence, in addition to the main displacements mentioned above, constantly undergo oscillations, which may be insignificant or important. The boundaries of the zones of uplift and subsidence are closely associated with these constantly changing oscillations: numerous local transgressions and regressions, unconformities, gaps in the deposition of sediments, and so on. Similar oscillatory changes in the zones of uplift and subsidence are examined in great detail in a work by V. V. Tikhomirov (1950) devoted to the Mesozoic history of the Lesser Caucasus.

Moreover, the region has undergone repeated general oscillations, producing temporary interruptions in the deposition of sediments and causing sharp regressions replaced by transgressions. Such oscillations occurred, for example, before the Late Jurassic, before the Late Cretaceous, before the Eocene, and a few times during the Tertiary. The individual oscillations are naturally reflected in the facies of deposits, establishing a more complex alteration of terrigenous and organogenic rocks than that which would have been produced by the basic movements alone.

The present-day structure of the Taurus-Caucasus geosyncline is represented diagrammatically at the top of Fig. 118. This structure is characterized by the alternation of large areas of uplift and equally large areas of subsidence. From north to south one sees the uplift of the Greater Caucasus, the Kura depression, the Rion lowland, the Trialet uplift, the uplift of the Somkhet mountains, the shallow basin of Lake Sevan, the uplift of the central Armenian highlands, the Araks depression, and the uplift of the Taurus range. It is not hard to see that these uplifts and depressions are not only the major elements of the present relief but also structural elements: the uplifts are enormous anticlines with outcrops of relatively older rocks at the surface, whereas the depressions on the whole have a synclinal structure and are filled with relatively younger rocks. For example, in the axial part of the Caucasus range Mesozoic and older rocks outcrop, but the Kura lowland is surfaced by Late Tertiary deposits. In the axial part of the Somkhet zone, Jurassic (and in places even older) rocks crop out at the surface, whereas there is an extensive development of Eocene and Upper Mesozoic rocks in the Sevan depression. In the central Armenian up-

land, under the very young lavas, thin Upper Mesozoic deposits and rocks of Early Mesozoic and Paleozoic age are found, but thick Tertiary deposits appear in the Araks depression.

Here we undoubtedly have structures that may be called synclinoria and anticlinoria. It was seen earlier that anticlinoria and synclinoria are formed as the result of the oscillatory movements of the earth's crust, its wavelike alternation of uplift and subsidence. It is also easy to establish the origin of inverted and noninverted anticlinoria and synclinoria. The structure of the Caucasus has already been analyzed from this point of view, and both inverted and noninverted structures have been distinguished within it (Chap. 6). The inverted anticlinoria are the main Caucasus range in its entirety, the Somkhet mountains, the Trialet range, and, outside the U.S.S.R., the Taurus mountains. All these anticlinoria were formed from the central uplifts that appeared within the corresponding intrageosynclines. Hence it is possible to understand the difference in the distribution of thicknesses and facies in the two complexes of deposits composing these anticlinoria. The lower complex dates from before the inversion, having been formed at the time when intrageosynclinal depressions existed on the sites of the anticlinoria. These deposits naturally form lenses, with their maximum thicknesses along the axes of the intrageosynclines. The upper complex of deposits was formed after the partial inversion, when the central uplifts appeared, and thus extend in the direction of the axes of the anticlinoria and thicken toward the adjacent depressions.

The anticlinorium in the central Armenian upland is not inverted. This is due to the fact that it is historically associated with a broad intrageanticline that was preserved as a stable area of relative uplift during the whole cycle. Naturally, here the thickness of all the deposits decreases toward the axis of the anticlinorium.

The problem of the origin of inverted and noninverted synclinoria is similarly solved. The Kura depression is an inverted synclinorium. An intrageanticline existed in its location during the Mesozoic. Thus the lower, or Mesozoic, complex of deposits is thinner at the bottom of the structural basin than in the neighboring zones; the thick upper, or Tertiary, complex, on the other hand, was formed at the time when the Kura depression, after a partial inversion, was transformed into an intermontane basin. The Rion depression is a noninverted synclinorium at whose center all the Mesozoic and Cenozoic deposits have their greatest thickness.

Development of Oscillatory Movements in Certain Other Geosynclines of the U.S.S.R.

None of the other geosynclines in the U.S.S.R. has been as thoroughly studied as the Caucasus. On the other geosynclines there are only frag-

mentary data on the history and development of the oscillatory movements in them and on the regularity of this development. Some of these data, however, are of great interest and will be briefly cited below.

Tien Shan. V. I. Popov (1938) and A. V. Peyve (1947, 1948) have devoted much attention to the history of oscillatory movements of the earth's crust in the Tien Shan mountains. The data obtained by these authors provide a general picture of the development of the northern Tien Shan (the Tien Shan ranges north of the Fergana Valley).

In the northern Tien Shan geosynclinal conditions existed, aside from Precambrian times, during the Caledonian and Hercynian cycles. At the beginning of the Caledonian cycle—in the Cambrian and the Ordovician—the maximum subsidence occurred in Talasskiy Alatau, where thick Cambrian arenaceous shale (flysch) deposits and Ordovician limestones were being accumulated. The latter are replaced to the north (in the Kirgiz range) by coarse clastic deposits, indicating that the subsidence of the Talass depression was bounded on the north by an eroded region of uplift. On the south this depression bordered on an uplift situated mainly in the area of the present Kuraminskiye Mountains and in the southwestern part of Chatkal'skiy range, where almost no Early Paleozoic sediments were deposited.

At the end of the Ordovician the conditions changed: in Talasskiy Alatau there was an uplift "which coincided with the zone of greatest subsidence in the preceding period of the Early Paleozoic" (Peyve, 1947). Thus at the beginning of the Caledonian cycle in the Talasskiy Alatau there was an intrageosyncline, which underwent partial inversion at the end of the Ordovician. The new central uplift was limited on the south by the deep Talass-Fergana fault, which occupies a prominent place in the structure of the northern Tien Shan. Along with the formation of the central uplift, marginal basins appeared on both sides, the southern one being the best developed. It is situated on the other side of the Talass-Fergana fault, in the area of the Ugamskiy, Pskemskiy, and Chatkal'skiy ranges. Here, in the Late Silurian and at the beginning of the Devonian, coarse clastic (molasse) deposits accumulated from the erosion of the Caledonian central uplift. But subsidence continued even beyond this time, and the Caledonian foredeep began to act as an intrageosyncline in the Hercynian cycle. Above the Silurian-Devonian molasse, limestones of Upper Devonian and Lower Carboniferous were deposited. During the Middle Carboniferous the subsidence changed to uplift (partial inversion of the Hercynian cycle), and a new zone of subsidence appeared even farther to the south, the Kuraminskiye range, where coarse clastic molasse deposits accumulated, as they also did on the site of the Fergana and Naryn Valleys. Thus the area of these valleys, up to the Middle Carboniferous, was an intrageanticline, and later became a typical intermontane basin.

In the southern Tien Shan a group of mountain ranges forms a large-scale tectonic complex: these are situated between the Fergana Valley to the north and the southern Tadzhik depression and the Alayskaya Valley to the south—the Nuratinskiy, Turkestan, Zeravshan, Karateginskiy, Alayskiy, and Gissarskiy mountain ranges. Here there was intensive subsidence of the earth's crust from the beginning of the Caledonian cycle to the second half of the Hercynian—from the Cambrian to the Early Carboniferous periods, inclusive. This was the time in which the thick sandstone-shale (flysch) strata of the Lower Paleozoic and the limestones of the Upper Devonian and Lower Carboniferous were deposited. This zone of subsidence (intraeosyncline) was bounded on the north by an uplift (intraeanticline) then existing on the site of the present Fergana and Naryn Valleys and on the south by a similar intraeanticlinal uplift in the area of the present southern Tadzhik depression, the Alayskaya Valley, the Peter I range, and the Zaalayskiy range. The existence of these uplifts is proved by the greater coarseness of the Lower and Middle Paleozoic deposits north and south of the axial part of this zone of mountains.

The inversion in the Zeravshan-Alayskiy intraeosyncline occurred before the Middle Carboniferous, when a central uplift (or a series of central uplifts separated by depressions) appeared, flanked by thick Middle and Upper Carboniferous flysch deposits. The central uplift continued to expand until the end of the Paleozoic, and an Upper Paleozoic coarse clastic molasse (Upper Carboniferous—Permian) was deposited in the Fergana-Naryn valley, in the southern Tadzhik depression, in the Alayskaya Valley, and in the area of the Peter I and Zaalayskiy ranges, where typical intermontane depressions were formed.

The Darvaz range farther south of the last intermontane basin has generally the same history as the Zeravshan-Alayskiy zone. Here, during the whole of the Caledonian and the first half of the Hercynian cycles, there was an intraeosyncline in which thick clastic deposits of the Lower Paleozoic and limestones of the Middle Paleozoic accumulated. After this an uplift appeared. To the north this was accompanied by the above-mentioned southern Tadzhik intermontane basin, and to the south, during the Late Paleozoic, a basin was formed in the southern Pamir, where molasse deposits of the Upper Paleozoic accumulated.

During the Mesozoic, typical platform conditions were established in the northern Tien Shan: the uplift of the ranges that appeared earlier and the subsidence of the depressions between them were retarded and did not exceed the rate and amplitude characteristic of the antecises and synecises on platforms. But along the southern margin of the Fergana Valley, during the Mesozoic and later, there were dislocations, with the formation of very large domical and boxlike uplifts, suggesting parageosynclinal conditions. These conditions are known to have ob-

tained in the area of the present Fergana range. Here during the Jurassic the subsidence of the earth's crust was much more intensive than in the other parts of the Fergana-Naryn basin. This region of subsidence was bordered by the deep Talass-Fergana fault on the west and extended along this fault from southeast to northwest. The basins were asymmetrical: the subsidence increased smoothly from west to east but was sharply limited on the east by the deep-seated fault mentioned above. At the beginning of the Cretaceous, there was a partial inversion here, as a result of which the Fergana range was elevated, dividing the Fergana-Naryn basin into two troughs: the Fergana and the Naryn. The area of the Fergana range is classed as a parageosyncline, because its structure is simple and has the character of a platform dislocation. Here also there is no igneous activity of a geosynclinal nature.

Parageosynclinal conditions at the beginning of the Alpine cycle have also been established in the Tien Shan area farther to the south. Thus the folding of Mesozoic and Cenozoic strata in the southern Tadzhik depression, the Peter I range, the Zaalayskiy range, and the southern Pamir is parageosynclinal.

The final changes in the structure of the Tien Shan are connected with a sudden revival of wavelike oscillatory movements in the middle of the Tertiary period. Uplifts and depressions formed at the end of the Caledonian and Hercynian cycles were leveled out to a certain degree during the Mesozoic, but they began again, to move respectively upward and downward with great contrast and intensity. The significance of this renewal of tectonic activity will be discussed later.

Thus the development of the Tien Shan clearly reveals the following regularities: dissection of the geosyncline into intrageosynclines and intrageanticlines, local inversions, and the formation of central uplifts and intermontane basins. After the geosynclinal development, a change to parageosynclinal development is noted. The new feature is the fact that some intrageosynclines show a development encompassing not one but two tectonic cycles: the Caledonian and the Hercynian (Fig. 120).

The Urals. The history of oscillatory movements in the southern Urals has not yet been studied. It can only be assumed that the axial uplift of the Urals—the Ural-Tau range—is a noninverted anticlinorium formed on the site of a stable intrageanticline that rose during the Caledonian and Hercynian cycles. The broad Zilair synclinorium farther to the west is filled mainly with thick Devonian volcanogenic strata and is apparently a parageosyncline. Consequently, this peripheral part of the Urals geosyncline is characterized by less intensive development. There are no reliable data on the interior parts of the geosyncline, on the eastern slopes of the Urals and under the young deposits in the interior depression of the Turgayskiy strait. The information available in the literature

on the stratigraphy and structure of these regions is in most cases out of date and needs to be reexamined.

In the central and northern Urals the few attempts to discover the regularity of the tectonic development refer to its eastern slopes, to the so-called greenstone zone. These include some very interesting investigations by A. V. Peyve (1947) in the northern part of this zone, in the vicinity of Serov, and by N. A. Shtreys (1951) in districts farther south: the Novo-Lyalinsk, Tagil'sk, and Kirovograd regions. According to his scheme of development, the greenstone zone in the Early Wenlockian age was a single depression in which thick volcanogenic rocks (porphyrite breccias) were accumulated. In the Middle and especially the Late Wenlockian age this area of subsidence, which possessed all the features of an intrageosyncline, was divided into three depressions by

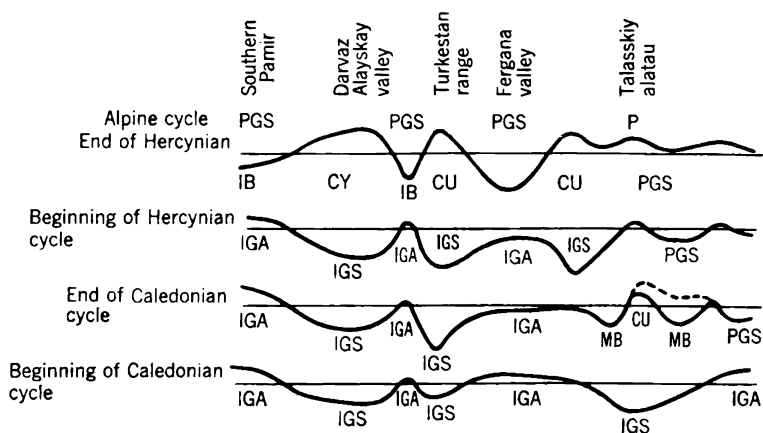


FIG. 120. Development of oscillatory movements of the earth's crust in Tien Shan.

the elevation along its axis of a narrow uplift and the preservation of two depressions at the sides. Later the uplift was sharply distinguished from the preserved zones of subsidence by its thin deposits, by the numerous gaps in deposition, and by its facies. At the time that the thick volcanogenic rocks were accumulated in the depressions, thin limestones were for the most part deposited on the uplifts.

Later a steady increase in the width of the central zone of uplift is observed, through a contraction of the zones of subsidence, which continued to the end of the Gotlandian age and even later. N. A. Shtreys noted that this widening of the central uplift took the form of a gradual accretion of newer and newer uplifts on both sides of the original anticlinal uplift (Fig. 121). In the older part of the central uplift at the end of the Gotlandian (during the Middle Ludlovian), there was an opposite movement: a rapid subsidence, with the formation of a superimposed trough of irregular shape, unconformably overlying the complex below. During this entire time the greenstone zone was bounded

on both the west and the east by very stable zones of uplift: the central Urals and the Isetsko-Saldinskiy uplift.

It is easy to see in this case the typical development of oscillatory movements in geosynclines that were mentioned above. In the intra-

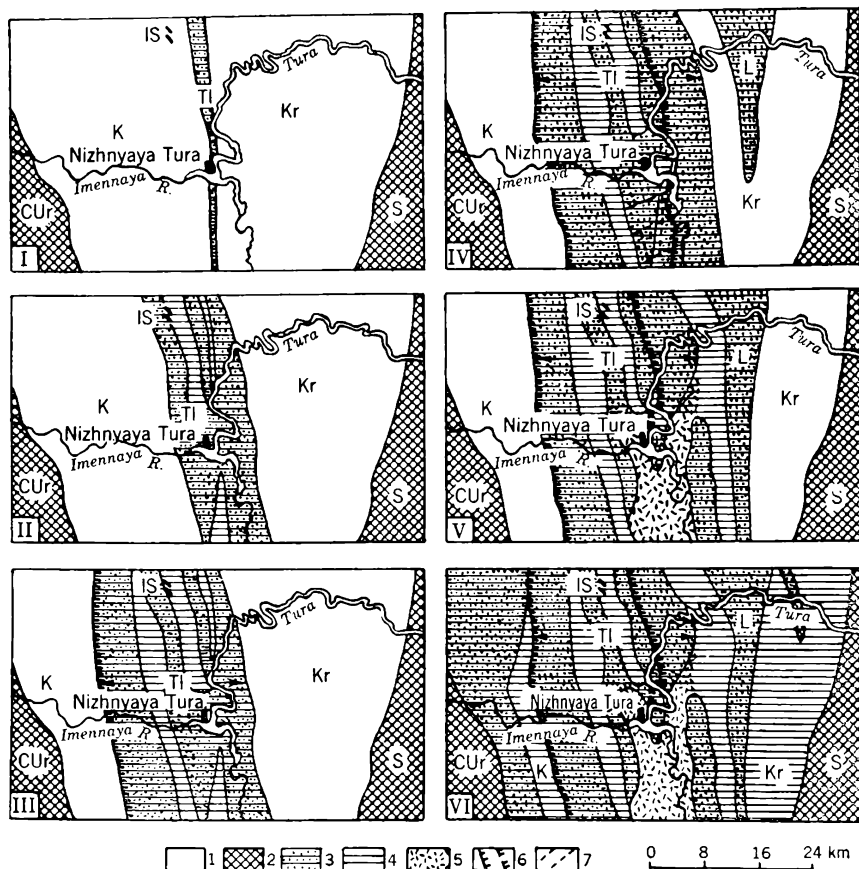


FIG. 121. Diagram of the development of the greenstone zone in the northern Urals in Gotlandian and Devonian times (after N. A. Shtreys): (I) end of Wenlockian; (II) middle of Lower Ludlovian; (III) end of Lower Ludlovian; (IV) end of Middle Ludlovian; (V) Upper Ludlovian–Downtonian; (VI) end of Eifelian; (1) marginal basins (K = Kuban, Kr = Krasnouralsk); (2) stable intrageanticlines (CUR = Central Urals, IS = Isetsko-Saldinskaya); (3) anticlines incorporated into the central uplift and formed successively along with the expansion of the latter; (4) synclines in the same position; (5) superimposed Tura trough; (6) deep-seated fractures; (7) faults (TI = Tagilsko-Isovskaaya anticline, L = Lyalinskaya anticline). The terminology of this diagram has been changed from that of its original author.

geosyncline of the greenstone zone, between two intrageanticlines, a central uplift rose during the Caledonian cycle, and depressions were preserved along its sides. The central uplift thereafter gradually widened.

Although a partial inversion is seen very clearly here, the process of development as a whole does not have the intensity that characterized the Caucasus intrageosynclines. The inversion did not end with a complete transformation of the intrageosyncline into an uplift. The adjacent intrageanticlines also maintained their proper regimes. Even with this less intensive development, however, the inversion is perfectly apparent.

Another new feature not found in the previous example is the formation of a superimposed trough. Such troughs were envisaged in the general scheme of geosynclinal development. N. A. Shtreys also noted a series of individual oscillations of the earth's crust that encompass both the entire region and certain local parts of it.

Central Kazakhstan and the Western Siberian Lowland. The Caledonian and the Hercynian geosyncline included a territory extending from the Urals in the west to the Yenisey River in the east. The folded region of central Kazakhstan belongs to this geosyncline and is its interior. The history of the tectonic development of this region is extremely interesting; nevertheless, it is nearly impossible to reproduce the course of this development because, in spite of the numerous investigations that have been made of the geologic structure of central Kazakhstan, there are not sufficient factual data for this purpose. To begin with, in order to trace the development of a geosyncline, one must have a summary of data on the distribution of the thicknesses and facies of the Paleozoic deposits. Up to now no such summary exists, and the available fragmentary information does not present a clear picture. This discussion will therefore be limited to very general observations (Shatskiy, 1938, 1940).

Fundamentally the geosynclinal development of central Kazakhstan ended with the Caledonian cycle. In the Hercynian cycle semiplatform parageosynclinal conditions prevailed here, and they were replaced in the Alpine cycle by typical platform conditions. The rocks of the Caledonian cycle are made up of two complexes. The lower complex comprises the Cambrian and the Lower Silurian and the upper one, the Upper Silurian and Lower Devonian. The Lower Cambrian is represented by quartzites, limestones, jaspers, jasper-quartzites, and various extrusive rocks from basic to intermediate in composition. The Middle and Upper Cambrian are composed of alternating conglomerates, sandstones, shales, and limestones, to which a layer of siliceous rocks has been added at the top. The Lower Silurian begins with basic extrusive rocks, tuffs, and tuffaceous sandstones, with interbeds of limestones and siliceous shales. Above these are intermediate and acid volcanogenic rocks with interbeds of limestones. At the end of the Silurian, gray-wackes appear in the section.

Only a very general regularity is to be noted in the distribution of the thicknesses of these rocks: the thicknesses to the west, in the zone of

the Ulutau range and the Kokchetavskiy massif, are considerably less (2.5 to 3 km) than in the east, in the region of the Chingiz range (7 to 9 km). Apparently, in the Caledonian cycle the Ulutau-Kokchetav zone was an intrageanticline, where one can see not only a decrease in the thickness of the Lower Paleozoic deposits but also the omission of a number of strata from the section.

At the beginning of the Late Silurian, numerous uplifts appeared within this extensive region of subsidence, which had the structures of anticlinoria. The Caledonian anticlinoria within central Kazakhstan are distinguished by the complex changes in their trends. In the west, near the Ulutau-Kokchetav intrageanticline, meridional trends are predominant, but to the east, in the area of the Chingiz range, the trends are northwest with a sharp turn to the northeast in the Bayanaul area. In the interior regions (in the northern Balkhash area) the Caledonian anticlinorium forms a sharp arc on the map, convex toward the north.

Upper Silurian deposits accumulated in the depressions between the Caledonian anticlinoria. The Llandoveryian stage is absent nearly everywhere, but where it exists, it is represented by coarse graywackes, tuffs, and porphyrites. Above lies a series of sandstones, shales, and conglomerates containing pebbles of more ancient rocks. The Upper Silurian wedges out in the direction of the Ulutau range and the Kokchetav massif, which were consequently a stable intrageanticline during the whole Caledonian cycle.

It is usually thought that the Caledonian anticlinoria of central Kazakhstan are central uplifts, formed on the sites of previous intra-geosynclines. Nevertheless, a final answer to this question is hard to find. Among the investigators who were studying the tectonic structure of central Kazakhstan in the nineteen thirties, the idea was commonly held that there was an "inheritance" in the development of the major structural forms in this region. "Inheritance" in this case was understood to have a very narrow meaning: it was supposed that the structural uplifts observed had been such during the entire course of known geologic history, as were the areas of subsidence (Shatskiy, 1938). The truth of this assumption can be determined only by tracing the changes in thicknesses from the uplifts to the depressions. But it is impossible to do this, because the Caledonian anticlinoria contain only the oldest rocks (Cambrian to Lower Silurian), whereas in the depressions only younger rocks are observed. The thickness of one and the same formation cannot be measured in both places.

Nevertheless, certain considerations cast doubt on the correctness of the concept of "inheritance" of structures under the given circumstances. It is known that the thickness of Lower Paleozoic rocks in the axial part of the Chingiz range reaches 9 to 10 km. In the adjacent Balkhash and Irtysh depressions the section through the Middle and part of the

Upper Paleozoic has approximately the same thickness. If, according to the theory of "inheritance," one considers that during the Lower Paleozoic the Chingiz range was a zone of relative uplift in which the thicknesses of the deposits must be relatively smaller, then in the depressions the Middle Paleozoic must overlie Lower Paleozoic rocks of a thickness considerably greater than 10 km. Therefore the total thickness of the Paleozoic in the depressions must approach 25 or 30 km. But such thicknesses are not known to exist anywhere on the earth and are apparently impossible. The only possible explanation of the enormous thickness of the Lower Paleozoic in the anticlinoria alongside the equally enormous thickness of the Middle and Upper Paleozoic in the depressions is that the Lower Paleozoic deposits are very thick only in the areas of present anticlinoria and are much less thick under the depressions filled with younger rocks. Thus one must suppose the existence, in the first part of the Caledonian cycle, of intrageosynclines on the sites of the present anticlinoria, which rose at the beginning of the Late Silurian, and the existence of intrageanticlines on the sites of the present depressions, which formed at the same time as the present anticlinoria. It is impossible to say whether these conclusions can be extended to all the Caledonian anticlinoria, but they seem sufficiently founded in relation to at least some anticlinoria and synclinoria, especially the anticlinorium of the Chingiz range and its adjacent synclinoria. This author believes that the Caledonian geosyncline of central Kazakhstan underwent a complex division into intrageosynclines and intrageanticlines and that there was a partial inversion.

This view is, in principle, close to the one expressed earlier in a treatise on the history of central Kazakhstan by the great expert on this region, N. G. Kassin (1937). Kassin distinguished mobile zones in this region ("geosynclines" and "mobile shelves") and "platforms" situated between them (Fig. 122). In the beginning great thicknesses of rocks were deposited in the "geosynclines," and the "platforms" were uplifted "blocks." Then the "geosynclines" were compressed into folds and uplifted, while the "platforms" became zones of subsidence. It is easy to see the present writer's intrageosynclines and intrageanticlines in N. G. Kassin's "geosynclines" and "platforms." On this problem A. D. Arkhangel'skiy (1947-1948) also agreed with N. G. Kassin: "It may be thought that . . . depressions are situated on the site of the Precambrian blocks, which had been transformed during the development of the geosynclinal region in the Hercynian period into areas of deep subsidence."

Middle Paleozoic deposits accumulated in the depressions between Caledonian anticlinoria. Over most of the territory of central Kazakhstan the Lower and Middle Devonian is represented by alternating basic, intermediate, and acidic lavas, tuffs, tuffaceous sandstones, and con-

glomerates, multicolored sandy argillites and cherts, and beds of limestone. Toward the Ulutau-Kokchetav uplift these rocks are replaced by a suite of red sandstones of no great thickness. In the Late Devonian, however, there was a redistribution and redistribution of the zones of uplift and subsidence over a considerable part of Kazakhstan. The Early Paleozoic tectonic zones trended primarily north-south; now a large region of central Kazakhstan was divided into narrow zones of

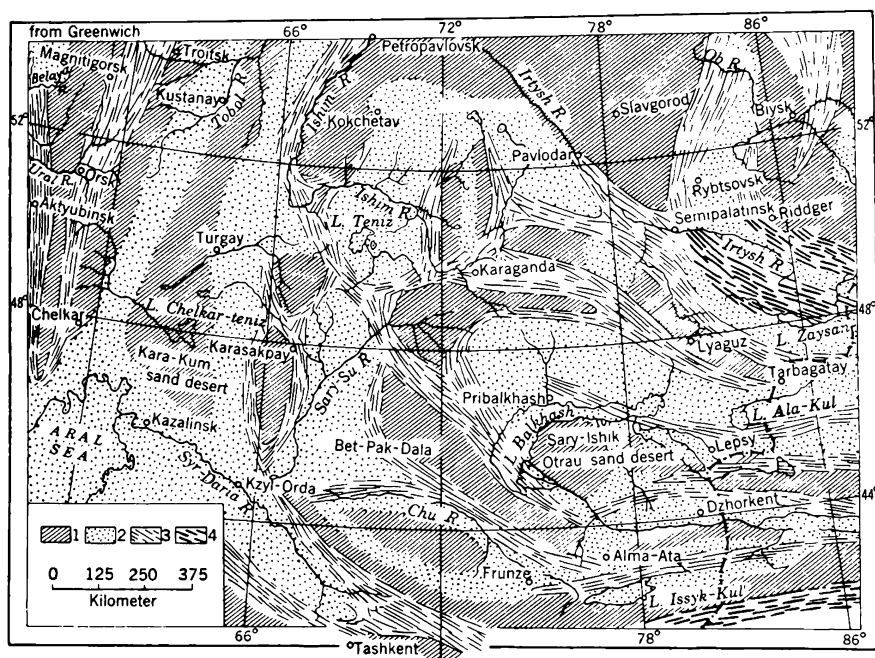


FIG. 122. Tectonic diagram of central Kazakhstan during the Middle and Late Paleozoic (after N. G. Kassim): (1) continent; (2) platform; (3) shelf; (4) geosyncline. In the author's terminology, (1) intrageanticline; (2 and 3) parageosynclines; (4) intrageosyncline.

uplift and subsidence trending northwest. These new uplifts and depressions closely resembled horsts and grabens in shape and were separated either by faults ("deep-seated breaks") or by steep flexures. The Upper Devonian and lower strata of the Carboniferous are represented mainly by limestones. Toward the end of the Early Carboniferous, coal-bearing sandy clay beds were deposited in the Karaganda and certain other depressions. Clastic rocks of Middle and Late Carboniferous and Permian age accumulated in a few places, including the Dzhezkazgan depression, the Tengiz trough, and part of the northern Balkhash area.

A large part of Upper Paleozoic depressions did not undergo inversion and was not subjected to intensive folding: here the dislocation of

the beds was limited to brachyantichinal uplifts of the discontinuous type, box folds, and narrow zones of small continuous folds along the faults that separated the area of subsidence from the uplift. But if the great amount of igneous activity, both extrusive and intrusive, that took place here is taken into account, a large part of central Kazakhstan during the Hercynian cycle may be considered a parageosyncline.

On the other hand, there are some regions of central Kazakhstan where especially thick Middle Paleozoic rocks were deposited (Devonian and Lower Carboniferous, up to 10 km) and where these rocks later underwent intensive folding of the continuous type. Such is the region south of the Tengiz depression (the Sarysu-Tengiz interstream divide), with its sharply defined Hercynian folds trending northwest, and some parts of the northern Balkhash area. Because of the lack of data, it is impossible to say whether Hercynian geosynclinal conditions developed in these regions, in the depressions that formed on the site of Caledonian intrageanticlines during the Caledonian inversion, or whether some intrageosynclines were formed at the beginning of the Caledonian cycle, continued to subside until the middle of the Hercynian cycle, and then were uplifted (their development running through two cycles, like the geosynclinal development of the southern Tien Shan).

In comparison with the preceding examples in central Kazakhstan, a new complex distribution of supposed intrageosynclines and intrageanticlines appears that differs sharply from that in the usual linear disposition of tectonic zones. Some depressions, formed as early as Hercynian time, continued to subside and to make room for the accumulation of sediments even later, in the Alpine cycle. Such are the Tengiz, Karaganda, and Maykyuben depressions, in the last two of which sandy clay coal-bearing Jurassic sediments were deposited. In the Tengiz depression the Mesozoic section includes continental deposits of Cretaceous and Tertiary age. These depressions during the Alpine cycle were in the nature of small platform syneclises.

North of the Kazakh folded region is the enormous western Siberian lowland, a vast basin filled with Mesozoic and Cenozoic deposits. The latter are represented almost entirely by sandy clay rocks of both marine and continental origin. In the Upper Cretaceous, diatomites and mold castings are encountered.

According to B. A. Petrushevskiy's survey (1951), the Mesozoic and Cenozoic deposits of the western Siberian lowland form three large depressions within it: the Chulym-Taz basin in the east, the eastern Urals basin in the west, and the Irtysh-Ob' basin between them (Figs. 123 and 124). The Chulym-Taz basin is bordered on three sides by uplifts of ancient rocks in the Yenisey range, the western Sayan, and the Kuznetskiy Alatau. On the west it is bordered by a zone of up-

lifted Paleozoic rocks (the Tom'-Kolyran folded arch) buried under younger sediments. The thickness of the Mesozoic deposits in central parts of the depression is apparently 1,200 to 1,500 m. Since the diameter of the depression is more than 100 km, the depression is very shallow and represents a typical platform syncline.

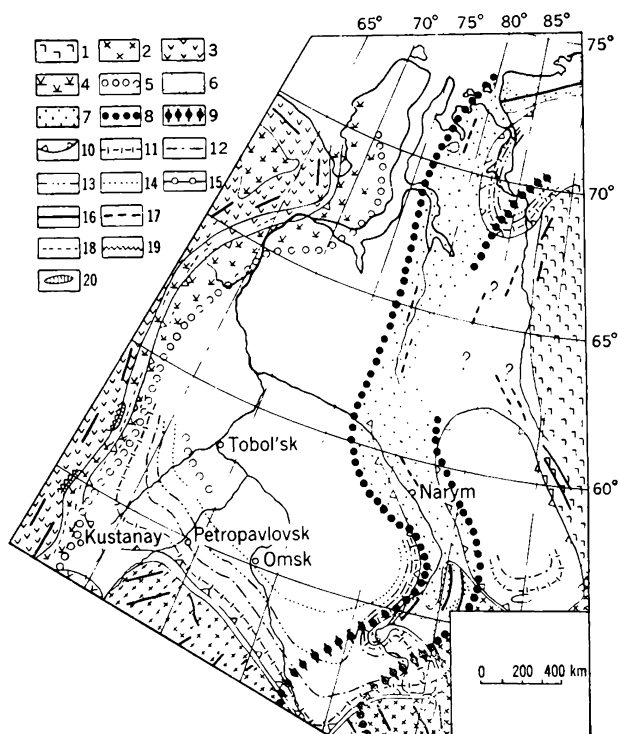


FIG. 123. Diagram of the structure of western Siberia: (1) regions of Precambrian folding; (2) regions with surface outcrops of Precambrian and Paleozoic folding; (3) regions with surface outcrops of Hercynian folding; (4) buried Hercynian structures; (5) boundaries of the areas of Hercynian folding; (6) regions with surface outcrops of late Hercynian folding; (7) buried late Hercynian structures (Ob'-Yenisey system); (8) boundaries of areas with late Hercynian structures; (9) the same, in areas of subsidence; (10) contours of synclises; (11) depth of pre-Mesozoic basement to 500 m from the surface; (12) the same, to 1,000 m; (13) the same, to 1,500 m; (14) the same, to 2,000 m; (15) the same, to 2,500 m or more; (16) main anticlinal structures in areas with surface outcrops of the basement; (17) supposed anticlinal structure of the Ob'-Yenisey system; (18) inherited anticlines in the zone of buried Hercynian structures in the Urals; (19) deep-seated fractures; (20) depressions associated with fractures (after B. A. Petrushevskiy).

The eastern Urals basin adjoins the Urals to the west and is the continuation of the Turgayskiy depression in the north, between the Urals and the Kazakh folded region. The eastern boundary of this basin is not completely clear: according to geologic and geophysical data, it is

marked by a buried northward continuation of the Kokchetav granite massif. The thickness of Mesozoic deposits in this depression does not exceed several hundred meters; this also indicates the platform nature of the tectonic conditions.

The center of the western Siberian lowland is occupied by the very large Irtysh-Ob' basin. The greatest thickness of the Mesozoic and Cenozoic deposits here exceeds 2,000 m and in places may reach 3,000 m. Toward the center of the basin the thicknesses of all the suites increase, and new strata appear that are missing from the section on the periphery of the basin. But the angle of dip of the beds produced by the subsidence of the basin in most cases is no more than 1° ; so that the basin is a platform syncline. Thus the western Siberian lowland must

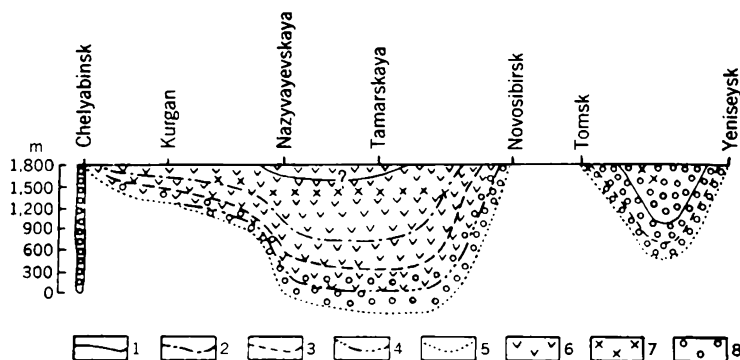


FIG. 124. Diagram of the subsidence of the southern parts of the Irtysh-Ob' and Chulym-Taz synclises: (1) position of the surface of the pre-Mesozoic formations toward the end of the Jurassic; (2) the same, toward the end of the Early Cretaceous; (3) the same, toward the end of the Late Cretaceous; (4) the same, toward the end of the Paleogene; (5) the same, toward the end of the Neogene; (6) marine facies; (7) lagoonal facies; (8) continental facies.

be considered as a part of the Alpine platform consisting of three large synclises separated by narrower antecises.

The structure of the Paleozoic basement of the lowland is of great interest. Thorough investigations of this problem were made by B. A. Petrushevskiy (1951), who believed that the structure of the Paleozoic under the western Siberian lowland has the following basic features: (1) in the west, along the Urals, under the lowland, stretches a buried Hercynian folded zone; (2) another buried Hercynian folded zone extends from the Rudnyy Altay to the northern tip of the Salair and beyond, along the middle course of the Ob' River and the entire course of the Pur River to the Taymyr range (this zone of folding appears in the Salair and in the Taymyr; in all other places it is buried under young deposits of the lowland); (3) the middle part of the lowland is situated on the northern extension of the structure of central Kazakhstan, which

was formed as the result of a geosynclinal development during the Caledonian cycle and mainly of a parageosynclinal development during the Hercynian.

Thus a rough sketch of the development of the Ural-Siberian geosyncline might be the following: During the Caledonian cycle there was an extensive geosyncline that occupied the area from the western slopes of the Urals to the Yenisey River. It was divided into numerous intrageosynclines and intrageanticlines with a linear disposition along the periphery of the geosyncline and a very complex, branching, and loop-shaped arrangement in its interior.

During the Hercynian cycle geosynclinal conditions were maintained to their fullest extent only on the periphery of the previous Caledonian geosyncline: in a zone along the eastern slopes of the Urals and in the Ob'-Yenisey zone mentioned above, which extends from the Rudnyy Altay to the Taymyr. In these two zones of the Hercynian geosyncline, intensive subsidence continued, ending in the Late Paleozoic with a considerable uplift. The Chulym-Taz syncline, like the Khatanga basin to the north, was a foredeep in relation to the Ob'-Yenisey Hercynian uplift.

The middle part of the Caledonian geosyncline—central Kazakhstan and the buried zone to the north of it—ended its geosynclinal development in the Caledonian cycle. During the Hercynian cycle parageosynclinal, and in some places even platform, conditions were established here. Geosynclinal conditions continued to exist only in certain limited areas. Moreover, the middle part of this territory did not undergo any great subsidence during the Hercynian cycle and, as a whole, in relation to the eastern Urals and the Ob'-Yenisey system, played the role of a vast and complex intrageanticline within the Hercynian geosyncline.

After the end of the Hercynian cycle, from the beginning of the Jurassic, an enormous interior depression appeared on the site of the previous geosyncline and completely covered the various elements of the geosyncline, leaving only individual fragments of it on the surface. Along with the formation of this depression, the following areas were involved in the subsidence: the basic Hercynian part of the Urals along the eastern slopes of the present Urals range; the whole northern part of the vast central intrageanticline; nearly all the Ob'-Yenisey Hercynian uplifts; and the eastern foredeep of the latter. In the east the interior depression directly adjoined the Siberian platform.

Southwestern Siberia. An exceptionally interesting, although tentative, analysis of the tectonic history of southwestern Siberia, including the western part of the eastern Sayan, the western Sayan, the Kuznetskiy Alatau, the Salair, the Altay, and the Gornaya Shoriya regions and the Tuviniian autonomous oblast' was presented by V. A. Kuznetsov (1948).

The geosynclinal development of this region took place mainly dur-

ing the Caledonian cycle. During the first part of the cycle, in the Cambrian and Early Silurian, different tectonic conditions prevailed in different parts of the region. In the western Sayan, on one side, and in a zone comprising the Gornyy Altay and the Salair, on the other side, great subsidence of the earth's crust occurred. In these two regions thick (on the order of several kilometers) strata of Cambrian and Early Silurian age were deposited—mainly sandy-argillaceous rocks with interbedded flows of basic lavas. On the other hand, in the western part of the eastern Sayan, in the area of the Minusinskaya basin, the Kuznetskiy Alatau, the Gornaya Shoriya, the eastern part of the Altay margin, and also along the other side of the western Sayan—in central Tuva—a very much abbreviated section of Cambrian rocks is observed, and the Lower Silurian is totally absent.

Thus in the first part of the Caledonian cycle this territory must have had two intrageosynclines and two intrageanticlines. These areas form arcs on the map that bend sharply toward the northwest. In the middle lies the central Tuvian intrageanticline. It is surrounded on the north, the west, and apparently the southwest (in the territory of Mongolia) by the western Sayan intrageosyncline. Farther outward a similar arc is formed by the intrageanticlines of the eastern Sayan, the Gornaya Shoriya, and the Anyuy-Chuya depression on the eastern slopes of the Gornyy Altay. Still farther out lies the Gornyy Altay and Salair intrageosyncline. On its outer edge the latter was bounded, according to certain data, by still another intrageanticline occupying the site of the present Rudnyy Altay (Fig. 125).

In the Late Silurian the conditions changed. Upper Silurian and younger deposits are absent in the regions distinguished above as being intrageanticlines. Thus these regions from the beginning of the Late Silurian began to be uplifted intensively (partial inversion). On the other hand, in central Tuva, where intrageanticlinal conditions had existed earlier, very thick series of sediments, including Upper Silurian, Devonian, Carboniferous, and Jurassic deposits, were deposited. The Upper Silurian is represented by coarse sandy clay deposits that are red-colored at the top.

Isolated depressions began to subside at this same time on the intrageanticlines of the eastern Sayan and Gornaya Shoriya and in the Anyuy-Chuya district. Such a depression was formed on the site of the present Minusinskaya basin. A narrow depression also appeared in the eastern foothills of the Gornyy Altay—in the Anyuy and Chuya river valleys. On the other side of the Altay, in the Rudnyy Altay and Kalba range, in the opposite foothills of the Gornyy Altay, a depression was also formed in which thick Middle and Upper Paleozoic deposits accumulated.

The Rudnyy Altay depression, one of the intermontane basins that

appeared at the end of the Caledonian cycle, changed during the Hercynian cycle into an intrageosyncline with renewed, fairly intensive folding. This intrageosyncline had its continuation in the Tom'-Kolyvan folded arch and farther on in the buried Ob'-Yenisey ridge mentioned above. Thus in the intrageosynclines of this extensive region there was

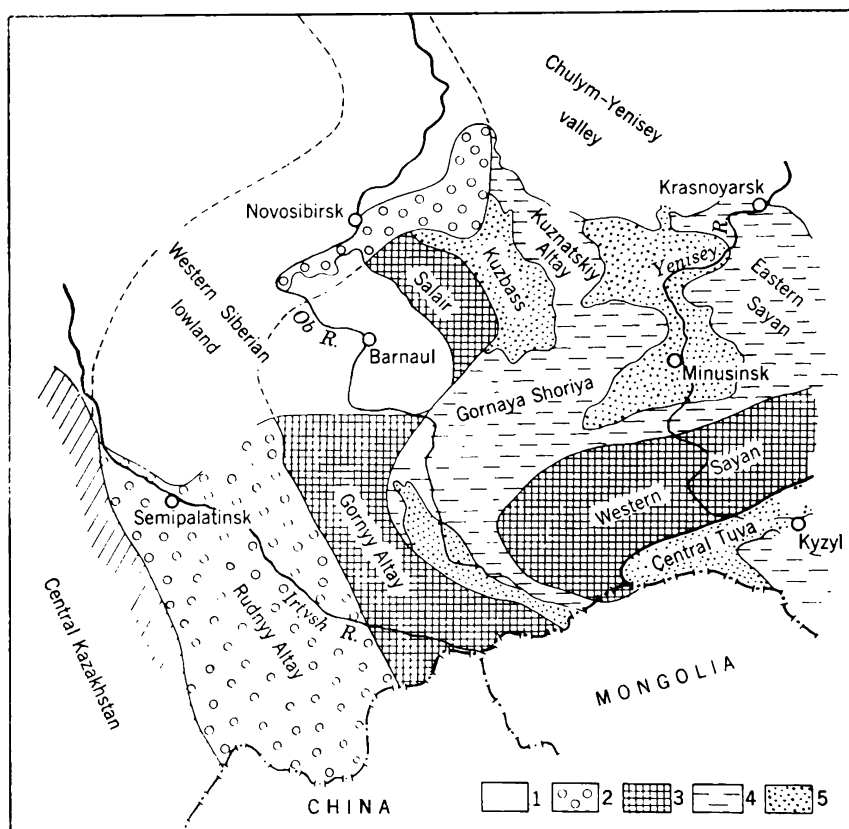


FIG. 125. Tectonic diagram of southwestern Siberia: (1) interior depression, filled with Mesozoic and Cenozoic deposits; (2) Hercynian intrageosyncline; (3) Caledonian intrageosynclines transformed into uplifts after inversion; (4) regions where Caledonian intrageanticlines maintain a steady rise; (5) regions where Caledonian intrageanticlines underwent subsidence after inversion (after V. A. Kuznetsov, as interpreted by V. V. Beloussov).

a partial inversion, and a central uplift arose; on the neighboring intrageanticlines intermontane basins appeared to compensate them—these either were single basins or were composed of individual depressions.

V. A. Kuznetsov also noted a number of secondary occurrences in the Caledonian history of this region. The most important of them, a general uplift of the region after the Middle Cambrian, coincided in time with the so-called Salair phase of folding. This upheaval continued

through the Late Cambrian, the deposits of which are everywhere absent. Subsidence, with the previous division of the area into intra-geosynclines and intrageanticlines, was renewed in the Early Silurian. Thus this uplift still belongs to the middle of the first, preinversion stage of development of the geosyncline, and it caused no great changes in the distribution of wavelike oscillatory movements.

Development of Oscillatory Movements in Certain Geosynclines Outside the U.S.S.R.

Two geosynclines may serve as examples: the Celtiberian geosyncline of the Alpine cycle in Spain, and the Caledonian geosyncline in England. The development of a number of other geosynclines in Europe and other parts of the world will be examined briefly later, in a survey of the present tectonic structure of the earth.

The Celtiberian Geosyncline. This geosyncline is located in Spain, between the valleys of the Ebro and the Tagus Rivers. Here during the Alpine cycle, in the Mesozoic and Cenozoic, oscillatory movements of the earth's crust were considerably more intense than in adjacent regions to the north and south. The Celtiberian geosyncline (Fig. 126) has the shape of a bay closed to the west. To the east it is connected directly to the great Mediterranean oval of the Alpine geosyncline. On the basis of structural investigations by Brinkmann (1931) and by

Richter and Teichmueller (1933), the following succession of movements took place in the Celtiberian geosyncline. Subsidence of the earth's crust began during the Early Triassic, as evidenced by the accumulation of sediments and the enlargement of the marine basin. North and south of the geosyncline at that time there were areas of uplift and dry land: the Aragon and Castile massifs, from which the clastic material was derived. Subsidence did not follow the same course everywhere: it was most pronounced at the margins of the geosyncline and least intensive in the center. Thus the geosyncline was divided into two intrageosynclines with an intrageanticline between them. The subsidence was compensated by the deposition of sediments, so that the sea remained shallow and the greatest thicknesses were accumulated in the intrageosynclines. The intrageanticline between them is called the

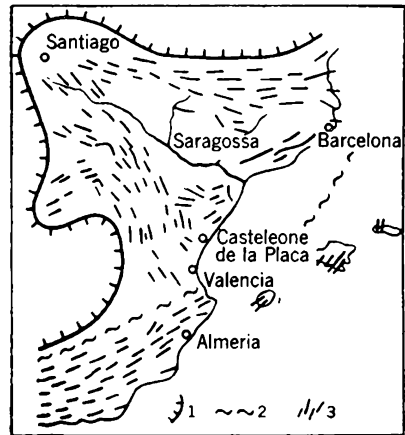


FIG. 126. Diagram of the Celtiberian geosyncline: (1) boundary of the geosyncline; (2) limits of intensive folding; (3) folds.

Ateca or *Montalban arch* by the authors mentioned above; it is distinguished by smaller thicknesses of the deposits, by coarser facies, and by many gaps in section, indicating that it often rose above the level of the sea.

During the Jurassic and Cretaceous periods, the movements of the earth's crust were distributed in the same manner: in the two intra-geosynclines intensive subsidence continued along with the accumulation of thick strata of sediments, whereas in the central intrageanticline an alternation of slight subsidence and uplift and erosion was to be observed. The internal division of the geosyncline is clearly seen in

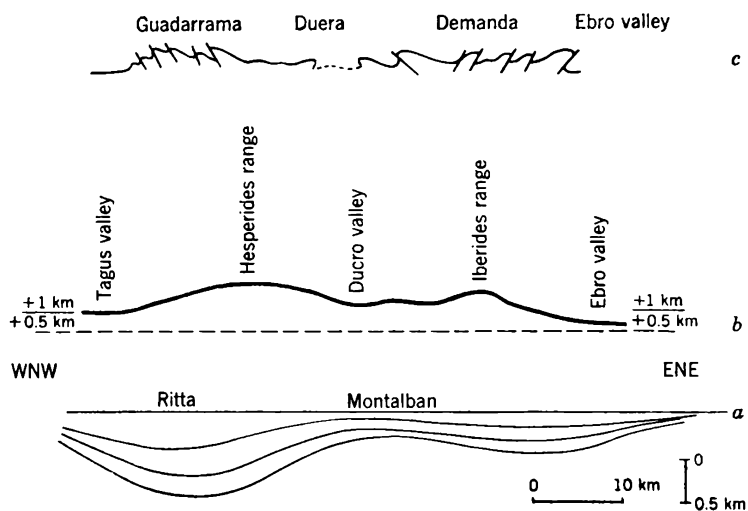


FIG. 127. Development of the Celtiberian geosyncline: (a) distribution of thicknesses of the lower Cretaceous; (b) Neogene uplift; (c) folded structures.

Fig. 127, which shows the distribution of thicknesses in the section through the Lower Cretaceous deposits of the geosyncline.

During the entire Mesozoic there was a gradual expansion of the geosyncline, especially toward the south, and regions that earlier had been outside its limits were involved in the subsidence. Moreover, the sea was expanding and transgressing upon the dry land. Against the background of this general trend in the development of the geosyncline, there were secondary movements complicating its history. Thus at the beginning of the Cretaceous, there was a temporary uplift of the regions bordering the geosyncline on the north and south, with a short-lived reduction of the area of the geosyncline and of the marine basin. This period is characterized by the wide development of continental facies. The transportation of clastic material into the geosyncline was intensified, and the slower rate of subsidence became insufficient to com-

pensate for this transportation, with the result that the surface of the deposits rose above the level of the sea. Before the Late Albian the intrageanticline was considerably uplifted for a short time; as a result, the Upper Albian in many places lies transgressively and unconformably on the Paleozoic, whereas in the intrageosynclines the change from the earlier deposits to the Upper Albian is continuous.

But the predominant movement of the first order during the whole of the Mesozoic was subsidence of the earth's crust. The distribution of facies indicates that the clastic material being deposited was derived mainly from the platforms to the north and south. On the whole, during this time, the intrageosyncline acquired about thirty-five hundred meters of sediments while the intrageanticline received no more than a few hundred meters.

During the Paleogene, a considerable transgression of the marine basin took place on the margins of the geosyncline, and the area of subsidence expanded greatly in both directions. In the north it embraced the whole Aragon massif and in the south a considerable part of the Castile massif. At the same time great changes were occurring within the geosyncline. These may be described in the words of the authors of the work from which this information is derived (Richter and Teichmueller, 1933):

The facies and thicknesses of the Paleogene deposits indicate that a total reversal of the trend of movement occurred during the Paleogene. The Celtiberian geosyncline which for so long had been a subsiding basin, for the first time became an area of erosion producing clastic material. The neighboring zones which had been elevated earlier were transformed into marginal zones of subsidence where the detritus from the rising "orogeny" accumulated in great thickness: the sinking Castilian shelf became the Tagus basin during the Paleogene, while the high Aragon massif became the Ebro basin. The old Ateca arch also behaved like the large massifs: although it did not sink completely, a depression appeared in its center.

Thus the distribution of movements changes sharply in the Celtiberian geosyncline between the Cretaceous and Tertiary periods. Where intrageosynclines once existed, there were now uplifts, the intrageanticline subsided considerably, and great depressions appeared on the periphery and included parts of the platforms.

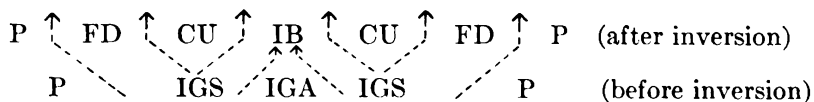
If one traces the subsequent history of the region, it will appear that the new uplifts thus formed expanded greatly in elevation and area, providing the sources of much clastic material, which was deposited in the rapidly sinking marginal basins. The Tertiary deposits in these basins are more than 1,000 m thick. With the growth of the uplifts, the marginal basins were displaced farther to the outside and their width decreased. The facies of the sediments in the basins changed from

marine to continental because the volume of the basins was smaller than that of the material being carried into them.

Finally, during the Late Pliocene, there was a general anticlinal uplift of the whole Celtiberian region. This uplift, however (see Fig. 127*b*), was differential, corresponding to the division of the geosyncline at the beginning of the Tertiary period: the greatest uplift occurred where there had been intrageosynclines during the Mesozoic, that is, in the areas of greatest subsidence. In the area of the former intrageanticline the uplift was minimal and even less on the periphery of the surrounding platforms.

Comparison of all these data leads to the conclusion that two stages may be distinguished in the history of the Celtiberian geosyncline. The first includes the entire Mesozoic and is characterized by the distribution of the interior zones described above and, for the most part, by the predominance of subsidence over uplift, as reflected in the accumulation of great thicknesses of sediments and in the gradual widening of the whole geosyncline and its corresponding marine basin. The second stage occurred during the Cenozoic and was characterized mainly by a predominance of uplift. Although subsidence at first continued along the margins of the geosyncline, the areas of subsidence gradually decreased and the process ended in a general uplift and in a retreat of the sea from the whole territory, with the formation of mountains. In addition, the distribution of regions of intensive uplift, on the one hand, and of regions of subsidence or of less uplift, on the other hand, was not the same in the second stage as in the first. More precisely, it was reversed: in the place of the original intrageosynclines, regions of great uplift were formed; but the intrageanticline was lowered as far as the edges of the adjacent platforms. Thus during the transition from the Mesozoic to the Cenozoic eras there was both a general and a partial inversion in this geosyncline.

The changes that took place in the Celtiberian geosyncline during its development may be represented schematically as follows:



where P = platform

FD = foredeep

CU = central upheaval

IB = intermontane basin

IGS = intrageosyncline

IGA = intrageanticline

arrows = direction of displacement of the margins of the respective zones

The Caledonian Geosyncline. Very extensive territories in Europe were characterized by geosynclinal conditions during the Caledonian cycle. England occupies only a small part of the Caledonian geosyncline. But this part is particularly suitable for study because after the Caledonian cycle, platform conditions were established in this area, so that the results of all the tectonic movements taking place during the Caledonian cycle are well preserved here.

The development of the Caledonian geosyncline in England begins in the Cambrian and ends in the Devonian.² Examination of the distribution of the Cambrian and Lower Silurian (Ordovician) shows that the English part of the geosyncline was divided into four intrageosynclines and three intrageanticlines located between them. In addition, there are also two intrageanticlines bordering this entire tectonic complex on the north and south (Fig. 128). The trend of all these zones is from southwest to northeast. From south to north we encounter (1) the Cornish IGA, (2) the Welsh IGS, (3) the Irish Sea IGA, (4) the Ettrick Bridge IGS, (5) the Moffat IGA, (6) the Girvan IGS, (7) the central valley IGA, (8) the northern Scotland IGS, and (9) the Louis IGA.

The intrageosynclines are distinguished by their great thicknesses of Cambrian and Ordovician deposits and by the completeness of their stratigraphic sections. The deposits are thin in the intrageanticlines. Comparison of thicknesses of deposits in one of the Ordovician stages (the Lower Bell) presents the following picture: thickness of these deposits in the Irish Sea IGA, zero; along the axis of the Welsh IGS, 1,800 m; farther to the southeast, approaching the Cornish IGA, the thickness decreases rapidly to 1,200 m, then to 600 m, and finally (in the Midlands) to zero. The thicknesses of other stages also change, and a similar picture can be observed in all the intrageosynclines.

On the margins of the intrageosynclines, and especially on those of the Welsh IGS, one can see clearly that the younger strata spread farther and farther, extending over regions where deposits did not occur before. Thus one observes a transgression of the sea from intrageosynclines to the intrageanticlines and an expansion of the regions of subsidence at the expense of the areas of uplift. As a result, the lower beds of the Ordovician extend beyond the Cambrian deposits and at the margins of the intrageosynclines lie unconformably on Precambrian rocks. The upper strata of the Ordovician spread even farther and also lie on the Precambrian. The Cambrian and Ordovician deposits are

² The history of the Caledonian geosyncline in England, based on the analysis of data from a number of English geologists, has been examined in detail by M. V. Gzovskiy jointly with the present author in another work (Belousov and Gzovskiy, 1945). This gives the sources from which the original factual material was taken. The work of O. Jones (1938) is of the greatest importance.

composed almost exclusively of terrigenous rocks (if volcanogenic rocks are not counted). The most common facies are those of the graptolite shales and graywackes. Only in the very uppermost Ordovician strata in the section do considerable thicknesses of limestone appear.

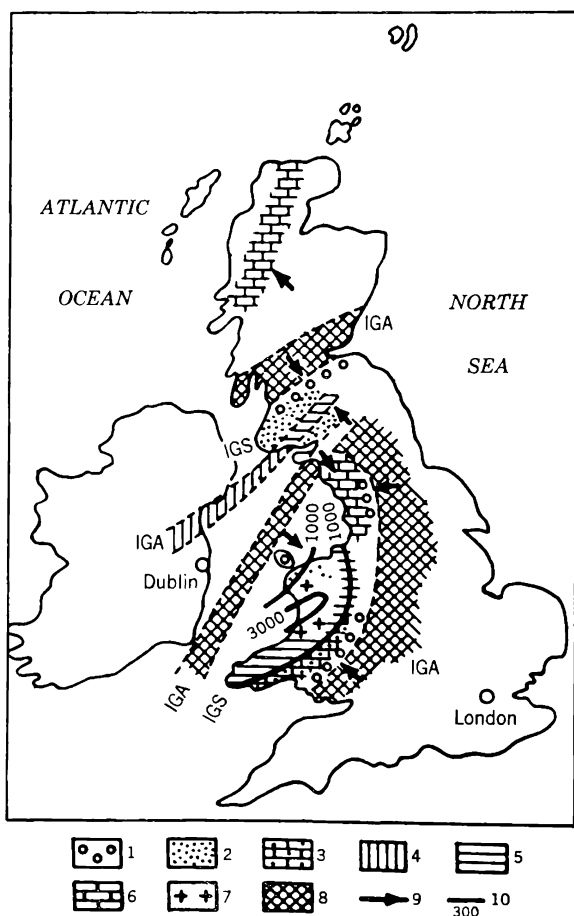


FIG. 128. Caledonian geosyncline of Great Britain during the Ordovician (Bell stage): (1) conglomerates; (2) sandstones and graywackes; (3) shell facies of calcareous sandstones; (4) shales; (5) graptolite shales; (6) limestones; (7) acidic volcanogenic rocks; (8) regions of erosion; (9) direction of transportation of clastic material; (10) thicknesses, in meters.

The facies distribution shows that the clastic material was carried in from the intrageanticlines. But not all the intrageanticlines were alike in this respect. The Irish Sea IGA supplied a great amount of clastic material in the direction in which the graptolite shales are replaced by graywackes in the Welsh IGS. The Cornish IGA produced much less

material. Because of these irregularities in the receipt of clastic material, a zone of very thin deposits of graptolite shales is displaced (in relation to the axis of maximum subsidence of the Welsh IGS) to the southeast, or in the direction from which the supply was the smallest. This displacement is of particular significance, because the attempt has often been made to draw the axis of subsidence along the zone of the thinnest deposits. This is incorrect: the axis of subsidence must be determined according to the thickness (Fig. 129).

There is need to dwell at greater length on the relationship between facies and thicknesses in northern England, in the region between the Irish Sea intrageanticline and that of the central valley. This discussion will be confined to the beds of the Bell stage (Upper Ordovician). In the central part of the southern highlands of Scotland (around Moffat), the Bell stage is represented by graptolite shales 70 m thick. Toward the northwest the shales are soon replaced by thicker graywackes. Still

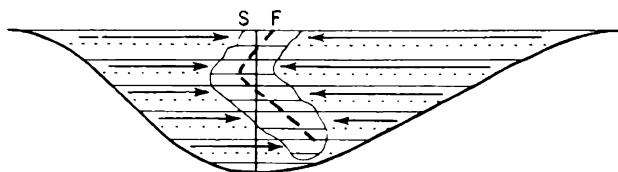


FIG. 129. Displacement of the facies axis of the intrageosyncline in relation to the axis of subsidence, as a function of the predominant direction of transportation of clastic material: *S*, axis of subsidence; *F*, facies axis, changing its position with time; horizontal arrows, direction of transportation of clastic material.

farther to the northwest the coarse sandstone deposits of the Bell stage are 1,100 m thick. Consequently, the material must have been transported from the northwest—from the central valley IGA. In the opposite direction from Moffat, toward the southeast, sandstones appear, interbedded among the graptolite shales; the thickness of the sandstones grows rapidly to 60 m at Ettrick Bridge. These sandstones indicate transportation of the material from the south—from the Irish Sea IGA. The thickness has also been observed to increase in that direction.

The English researcher Jones (1938), in studying the development of the Caledonian geosyncline, stressed the association of the finest facies in the Moffat area, which he took to be the axial part of the "Moffat geosyncline." He explained the decrease in the thicknesses of the deposits in the Moffat zone by the fact that the clastic material did not reach the central, and consequently the deepest-water, parts of the geosyncline.

Nevertheless, Jones repeatedly stressed the fact that the graptolite shales could not be considered deep-water deposits. This agrees completely with the opinion of most present-day authorities and with the

material presented in the famous summary by Twenhofel (1936), who considered the graptolite shales to be analogous to the black ooze deposited on the bottom of some modern very shallow seas (such as the Baltic Sea). But even if it is allowed that the depth of formation of the graptolite shales, as indicated by S. Bubnoff, actually amounts to 500 m, from the tectonic standpoint this figure is greatly exaggerated, whereas the Girvan rocks, whose depth of deposition is considerably less, will be considered to have been formed at the surface of the water. If to these depths one adds the thickness of the strata deposited during the Bell age, then the base of these deposits in the Girvan will be 500 m lower than in the Moffat.

From this it is clear that the Moffat zone during the Bell age was a region of less subsidence than the areas farther to the north (Girvan) and to the south (Ettrick Bridge end): that is, it was an intrageanticline.

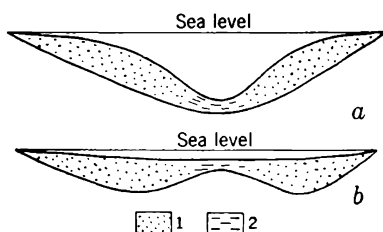


FIG. 130. Accumulation of deposits on the bottom of the sea (*a*) in the overwhelming majority of cases, an irregular diagram; (*b*) usually regular diagram; (1) sands; (2) clays.

diagrams of the accumulation of deposits on the bottom of the sea as represented in Fig. 130*a*. Such a diagram is nearly always misleading, since it is based on a completely erroneous conception of an immobile marine basin that is gradually filled with deposits.

It has been emphasized that deposition occurred simultaneously with the subsidence of the crust and that these two processes, as a rule, are in balance. If the thicknesses of the deposits in diagrams *a* and *b* (Fig. 130) differ by more than 100 to 200 m, then almost certainly the small thickness is determined not by transportation of a small amount of material into a large basin but by less accumulation of deposits because of the small subsidence (Fig. 130*b*).

In studying such instances, one must establish as accurately as possible the depth of formation of deposits and, in relation to the distribution of the depths, construct a profile of the bottom of the basin. Using this as a datum plane, one must then plot beneath it the thicknesses of deposits accumulated during the interval of time under consideration. The lower curve thus obtained will indicate the actual distribution of

Its geanticlinal character was manifested in the fact that it subsided more slowly than the surrounding areas. Therefore it was not an area of erosion, and thin terrigenous sediments, which were brought in from the opposite margins of the adjacent intrageosynclines, could accumulate here. This zone was farthest away from the areas of erosion, and consequently the thinnest deposits are observed in it. This case is of significance in principle. Textbooks show

the earth's crust. In any case, one should not use the diagram shown in Fig. 130a to determine the depth of the basin, as recommended in some textbooks. For example, after determining that at the margin of the ancient basin the thickness of the deposits is 1,000 m and at the middle of the basin no more than 200 m, one will conclude that the depth in the middle of the basin is 800 m. In reality, the depth at the center of the basin cannot be much greater than the depth at its margins, and the smaller thicknesses will be the result of less subsidence of the bottom in this part of the basin.

Let us turn to the Upper Silurian (Gotlandian) in England. In Wales the facies and thicknesses of the Lower and Middle Llandoveryan stage are distributed as before and indicate the existence of the same intrageosyncline. But beginning with the Upper Llandoveryan, the conditions change radically. In the first place, during the Late Llandoveryan, Wenlockian, and Ludlovian ages, the Welsh IGS expanded to the east and northeast and new regions were drawn into the subsidence that before had sunk but little (for example, the Lake Country). The second and most important feature is that the distribution of the facies now indicates the existence of a rapidly eroding area of dry land in the central part of Wales, where the greatest subsidence and deposition had previously been. In approaching these regions, one also observes an increase in the coarseness of the deposits.

All this indicates that on the border line between Middle and Late Llandoveryan ages in the Welsh IGS there was a regional inversion of geotectonic conditions, with the formation of a central uplift. This uplift was bordered by zones of continuing subsidence of the earth's crust not only on the southeast and northwest (that is, across the trend of the whole intrageosyncline) but also along its trend, on the northeast: in the Lake Country subsidence and deposition continued during nearly the whole of the Gotlandian age, although to less degree than in other regions (Fig. 131). Thus subsidence encircled the central uplift like a horseshoe. It can be seen from the preceding diagram (see Fig. 128) that the Welsh IGS was closed along its trend to the northeast, becoming an intrageanticlinal region. Comparison of these facts shows that the central uplift occurred in the central, earlier-subsiding part of the intrageosyncline and gave way to subsidence toward its margins. The historical shape of the uplift after the inversion corresponds to that of the area of subsidence before inversion. Marginal subsidence of the Welsh IGS was to a considerable extent superimposed on the surrounding intrageanticlines, inasmuch as the whole IGS was expanding in area. The intrageanticlines were not destroyed during this process, however, but merely contracted accordingly.

The distribution of the facies and thicknesses in the northern regions suggests that, before the Late Llandoveryan, a regional inversion oc-

curred in the Ettrick Bridge intrageosyncline. In its interior a central uplift appeared, accompanied by marginal zones of subsidence. But since the tectonic zones in this part of England are narrow, these depressions were superimposed on the adjacent intrageanticlines. The zone of subsidence paralleling the Ettrick Bridge central uplift in the north, after "rolling" upon the Moffat intrageanticline from the south, embraced it completely, so that, instead of intrageanticlinal

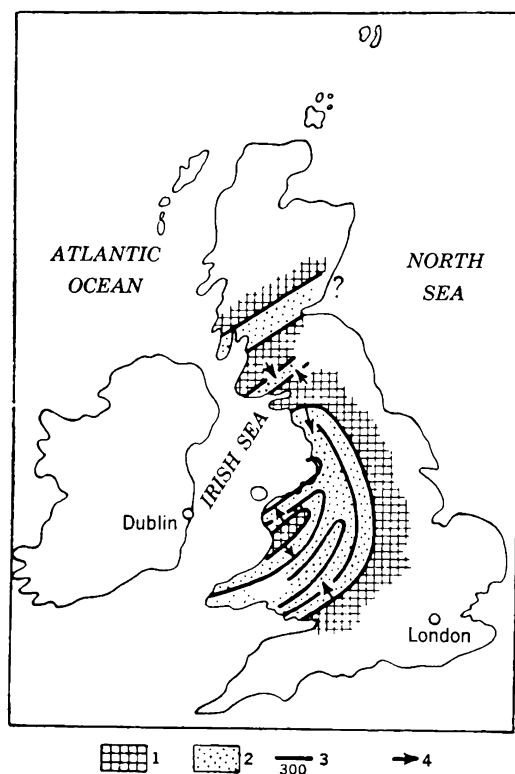


FIG. 131. Great Britain in the Caledonian cycle immediately after the inversion (Wenlockian): (1) areas of erosion; (2) zones of deposition; (3) isopachs, in meters; (4) direction of transportation of clastic material.

conditions, here one observes very intensive subsidence with a thick accumulation of clastic rocks carried in from the Ettrick Bridge uplift. The Moffat intrageanticline of the first generation was transformed into an intermontane basin.

During the entire Gotlandian, beginning with the Late Llandoveryan, the central uplifts in this area expanded, while the facies in the zones of continuing subsidence became coarser and changed from marine to continental. This testifies to growing predominance of uplift at the

expense of subsidence. Thus there are signs not only of partial but also of general inversion.

The last stage of development of the geosyncline took place during the Devonian period. The Devonian deposits in England are represented mainly by continental red beds and only to the south by marine deposits. During the Devonian period, England was divided, as before,

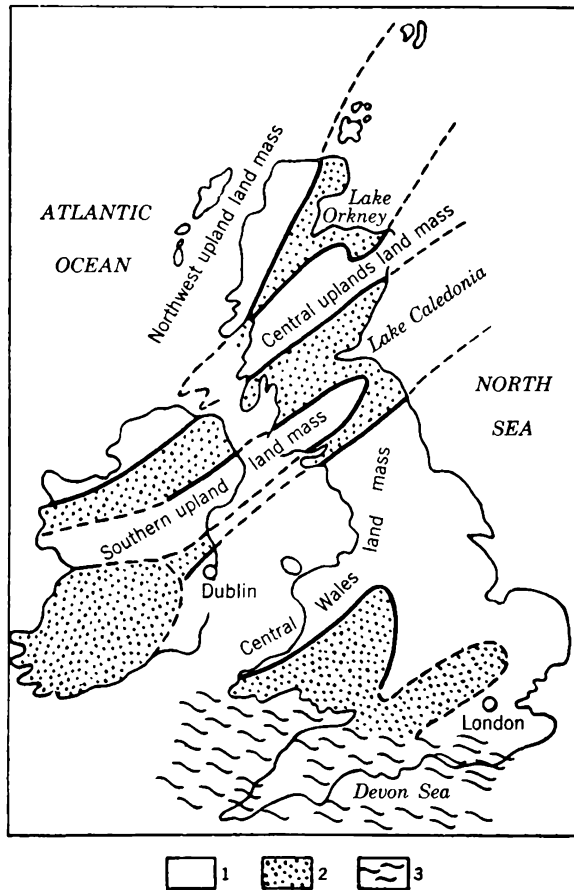


FIG. 132. Great Britain in the Devonian period: (1) areas of erosion; (2) continental deposits; (3) marine deposits.

into areas of uplift and erosion, on the one hand, and of subsidence and deposition, on the other. Both uplift and subsidence were very intensive during that period, so that England was a highly dissected mountainous country where uplift generally predominated over subsidence. Comparing the distribution of areas of accumulation of Devonian deposits with the previous disposition of intrageosynclines and intrageanticlines (see Figs. 128 and 132), we discover that during the Devonian the

subsidence and deposition occurred exactly on the sites of the pre-inversion intrageanticlines and that the rapidly rising areas of erosion were situated on the preinversion intrageosynclines. Thus the history of the Caledonian geosyncline in England shows the same regularities as do the preceding examples: separation into intrageosynclines and intra-

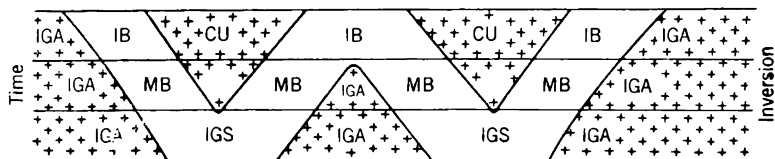


FIG. 133. Development of the Caledonian geosyncline in Great Britain.

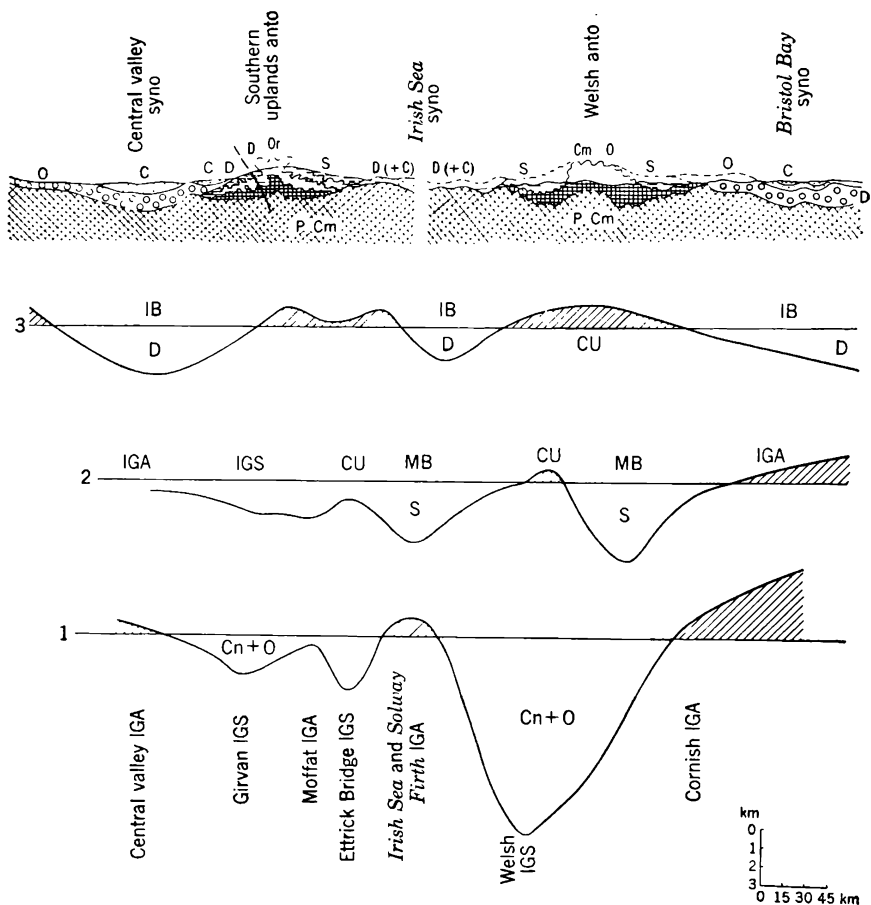


FIG. 134. Sections through Great Britain (top) and diagram of the development of oscillatory movements during the Caledonian cycle: (1) Cambrian and Ordovician thicknesses; (2) Gotlandian thicknesses; (3) Devonian thicknesses; areas of uplift are crosshatched.

geanticlines and predominance of subsidence at first, and then of uplift after partial and general inversions.

Diagrams of the development of this geosyncline are shown in Figs. 133 and 134. The upper part of Fig. 134 is a diagrammatic structural section through England. Since the structures here are inverted, the beds dating from before the inversion form lenses under the anticlinoria and the beds dating from after the inversion form lenses under the synclinoria.

Some Conclusions on the Regularities in the Development of Geosynclines

The examples described above are sufficient for a preliminary survey of the development of oscillatory movements in geosynclines. Some of the conclusions may be discussed in greater detail.

1. Geosynclines are divided into zones of higher-order intrageosynclines and intrageanticlines.

2. Two stages in the development of a geosyncline have been distinguished. The first is characterized by the predominance of subsidence over uplift, as reflected in the intensive subsidence of intrageosynclines. Transgression of the marine basin is associated with this progress. The second stage is characterized by a predominance of uplift, manifested in intensive elevation of uplifts and their expansion, in regression of the seas, and finally in formation of a mountainous relief. Consequently, in the first stage of development one may consider the geosyncline, to a first approximation, as a zone of subsidence and in the second stage as a zone of uplift. The change from the first stage to the second is a general inversion of the geotectonic conditions. The cycle terminates with a period of mountain building.

3. In the sense of a change in the regime of oscillatory, wavelike movements during the cycle of their development, each intrageosyncline behaves like the geosyncline as a whole. From an area of subsidence the intrageosyncline changes to an area of uplift, representing a partial inversion of the geotectonic conditions. But in all intrageosynclines the partial inversions occur simultaneously; as a result of the sum total of partial inversions, the intrageosynclines that existed before inversion become central uplifts, and zones of crustal subsidence are formed at their margins. The peripheral zones, which border the platforms on either side of the geosyncline, are called *foredeeps*. The new depressions formed at the margins of the central uplifts in the interior of the geosyncline are displaced toward the intrageanticline as the uplifts widen. If the intrageanticline is narrow, the two depressions moving upon it from both sides combine very rapidly and completely destroy the intrageanticline, changing it into a zone of intensive subsidence: an intermontane basin. Thus the disposition of the zones of uplift and

subsidence in the geosyncline after the inversion is the reverse of that before the inversion. When the intrageanticline is of considerable width, the basins after the inversion may lie mainly on the margins of this zone and its central part may preserve its intrageanticlinal character to the end of the cycle.

4. Some zones of subsidence with an intermediate, incomplete development are encountered in geosynclines. This incompleteness is reflected in the lack of a partial inversion. Such depressions are called *parageosynclines*. They are preserved during the whole cycle and are only partially drawn into the uplift or subsidence that approaches in a wavelike motion from the adjacent more active zones.

5. Along with the uplift and subsidence in geosynclines, large structural forms—anticlinoria and synclinoria—occur in the earth's crust in the form of "congealed" waves. The difference in thickness between the preinversion and postinversion strata in inverted structural elements of the geosyncline sometimes causes a lack of correspondence between structures in shallow and deep beds. For example, the increase in the thickness of the preinversion strata along the axis of the central uplift, where the greatest subsidence took place before the inversion, may cause the anticlinorium in shallow beds to change to a synclinorium in the lower strata. In exactly the same manner, intermontane basins are always synclinoria in the postinversion series but may be anticlinoria in the preinversion series, depending on the extent of the uplift in that location before the inversion. In parageosynclines and regions of preserved intrageanticlines, noninverted synclinoria and anticlinoria are formed whose structure is similar at all levels.

6. On some central uplifts superimposed troughs appear at the end of the cycle, during the period of mountain building, along with great interior depressions over the geosyncline as a whole. Both forms have a platform type of development.

All the transformations in geosynclines must be seen not only in cross section through the geosyncline but in the horizontal surface. In Western Europe there is a geosyncline that is composed of many individual open ovals. This form is typical of Alpine geosynclines in general. Much literature has been devoted to the so-called syntaxes, or sharp contraction of the width of the geosynclinal zone—the necks observed in some places. A particularly well-known neck is the one along the Pamir-Punjab line, where the Alpine geosyncline is greatly decreased in width. In connection with this phenomenon the concept has been created of *rigid* excrescences of the continents compressing the geosyncline. Many other mechanical constructions have been attempted. In reality, such contractions in the width of the geosyncline are observed where two adjacent geosynclinal ovals join. In the case of the Punjab neck, this connects the ovals of Iran and Tibet (Fig. 135). There was no

"squeezing" of the geosyncline here: from the very beginning (from the time the original region of subsidence began to develop) it had a beadlike shape.

It must also be remembered that closing of the individual intrageosynclines along their trend leads to a similar closing and subsidence along the trend of the individual central uplifts and anticlinoria. Because of the irregular disposition of the separate intrageosynclines within the geosyncline, the anticlinoria and synclinoria may be situated very differently. An en echelon disposition of anticlinoria and synclinoria is extremely characteristic. This is especially true of anticlinoria and synclinoria of secondary orders; the arrangement of these structural forms within the limits of the Somkhet meganticlinorium (see Fig. 65) may serve as an example.

The irregularity of the distribution of areas of uplifts and areas of subsidence in geosynclines hinders their unification into continuous, extended zones. Nevertheless, one often encounters attempts to mark out such continuous zones of anticlinoria and synclinoria, as in the case of the whole Alpine geosyncline of Europe and western Asia together. Such attempts are very artificial. One must never lose sight of the possibility of disconnected individual intrageosynclines and intrageanticlines within the geosyncline, or their joining along the trend. For a long time investigators have been trying to decide which ranges are the southeastern continuation of the Caucasus: the Kopet-Dag, the Balkhan, or the Elburz. But the Caucasus may itself be a completely closed system of two intrageosynclines with blind ends both in the southeast—on the Caspian shore—and in the northwest—in the Taman peninsula.

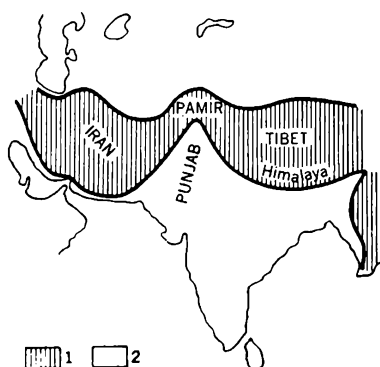


FIG. 135. Narrowing of the Punjab as a result of the joining of two geosynclinal ovals: (1) geosyncline; (2) platforms.

The Degree of Synchronization in the Development of Geosynclines

From what has been said above, it may be seen that during each geotectonic cycle there was a certain complex of geosynclines, which developed according to a similar sequence of events. The question arises whether the individual stages occur simultaneously in the development of different geosynclines.

The difference in time of the partial inversions in different intrageosynclines within the same geosyncline has already been mentioned. By comparing different geosynclines of the same cycle, located at great

distances from each other, one can also establish the differences in the times of occurrence of the various stages of the cycle. For example, the time at which the subsidence begins in the geosynclines of the Alpine cycle varies, in different cases, from the Permian period to the Middle Cretaceous. In most cases, however, the subsidence begins during the Triassic or the Jurassic. The subsidence of the Hercynian geosynclines began at various times from the Late Silurian to the Late Devonian.

Similarly, there are differences in the time of the inversion: in the Caledonian cycle the inversion, in most cases, occurs at the end of the Ordovician and the beginning of the Gotlandian. But in the northern Appalachians it apparently took place as late as the beginning of the Devonian, and in some geosynclines of southern Siberia it may have been as early as the Cambrian. There is a closer correspondence in the Hercynian cycle. The inversion nearly everywhere occurs at the end of the Early and beginning of the Middle Carboniferous, although local deviations from this time have been observed.

In the Alpine cycle the variations in the time of inversion are much greater. Geosynclines are known in which the inversion occurred during the Jurassic: one such is the intrageosyncline of the Sierra Nevada in the U.S.A. In the Somkhet intrageosyncline in the Caucasus, central uplifts in some places also appeared as early as the Late Jurassic. In many cases the inversion in the Alpine geosynclines happened at the end of the Jurassic and the beginning of the Cretaceous. An example of this is the Alpine geosyncline of East Asia (with the exception of the marginal zone of the geosyncline next to the Pacific Ocean, where the inversion took place later). In the Eastern Alps the inversion also occurred before the Late Cretaceous. In other European geosynclines of the Alpine cycle, the inversion occurred before the Eocene, as in the Western Alps, the Apennines, the Pyrenees, and other places. Finally, in the greater Caucasus the inversion took place at the end of the Paleogene.

These cases of differences in the time of each stage of development in particular geosynclines do not, however, violate the general periodicity and, in its general features, the synchronization of the process as a whole. In the first place, large deviations are observed in individual cases, but each cycle has its "preferred" time for the occurrence of particular stages. This time for the inversion is, in the Caledonian cycle, the beginning of the Gotlandian; in the Hercynian cycle, the end of the Early Carboniferous; and in the Alpine cycle, the beginning of the Paleogene. Secondly, in spite of certain differences in the time of particular stages, each geosyncline nevertheless runs through its development within the time interval of the cycle; and if, for example, one geosyncline reaches the stage of inversion earlier than another, this does not mean that the new cycle will begin earlier in it, but only that

the second part of the cycle, and especially the epoch of mountain building, will continue longer in this case up to the end of the given cycle.

It is appropriate here to raise the question of the so-called Mesozoic, or Pacific Ocean, cycle, mentioned earlier. The examples of the development of the movements in this cycle are encountered in many regions of eastern Asia (China, the Transbaikal, the Far East, northeastern Asia) and on the Pacific coast of North America. But from a geotectonic standpoint there is no basis for distinguishing a special Mesozoic cycle. The main feature of all these regions is only a very early inversion (occurring well before the end of the Jurassic) and an early beginning of mountain building. But the epoch of mountain building here lasts longer than in other places and continues until the end of the general cycle. Thus the entire complex of movements forming the cycle is "contained" within the usual frame of the cycle.

The Transition from One Cycle to Another in the History of Geosynclines

The discussion thus far has been concerned with the regularities in the development of geosynclines during a single geotectonic cycle. One can speak of Caledonian, Hercynian, or Alpine geosynclines or of the geosynclines of a given particular cycle, because in each cycle there is a periodic repetition of similar processes. On the other hand, we know that the periodicity of tectonic movements is only one aspect of the development of geosynclines. The other comprises the directional, linear changes continuing through all the cycles and uniting them into a single common succession of events. This is the time to discuss this aspect of the development of geosynclines, and the existence and forms of the transition between geosynclines of different cycles.

Observation shows that many geosynclines of earlier cycles have lost their mobility, become platforms, and preserved this state until the present. The Caledonian geosyncline of England became a platform at the beginning of the Hercynian cycle and remained one during the Alpine cycle. The Hercynian Ural-Siberian, Appalachian, and Western European (Spain, France, central Germany) geosynclines became regions of platform development from the beginning of the Alpine cycle. There were Precambrian geosynclines undoubtedly in every place in which, from the Cambrian (from the beginning of the Caledonian cycle), platform conditions existed and geosynclinal movements no longer took place. The Russian plain is an example.

On the other hand, almost everywhere that a geosyncline of one cycle or another is encountered, one may discover signs of the existence of geosynclinal conditions during previous cycles as well. The Cordilleras of North America are especially interesting from this standpoint: this region was a geosyncline during the Alpine cycle. But the great thicknesses of the Paleozoic deposits and the nature of the movements during

the Paleozoic era show that a geosyncline of the Hercynian cycle existed there and, before that, a geosyncline of the Caledonian cycle.

In spite of the small development of Paleozoic deposits in the Alps, it can be established that geosynclinal conditions obtained here also, not only during Alpine but also during Hercynian and Caledonian times, without mentioning the Precambrian. Clear indications of geosynclinal development during the Hercynian and Caledonian cycles are found in the area of the Alpine Taurus-Caucasus geosyncline. The Caledonian cycle of geosynclinal development is also very clearly reflected in the vicinity of the Hercynian Ural-Siberian geosyncline. Many such examples can be given, all of them leading to the conclusion that geosynclines of succeeding cycles are regularly developed where geosynclines existed in previous cycles. Inasmuch as later geosynclines develop on the location of earlier ones but, on the other hand, many geosynclinal regions of earlier cycles later become platforms, one can confirm that the area of the earth's surface occupied by geosynclines with the passing of time (in the transition from one cycle to another) becomes smaller and the territory occupied by platforms increases.

This regularity is manifested very clearly, for example, in Europe. The Caledonian geosyncline in Western Europe was very wide. Its southern margin is found in the central Sahara, and Spain, France, Belgium, and England are contained within its limits. Its northern margin is unknown, because the geosyncline extends northward from England under the Atlantic Ocean. It is possible that this margin should be sought as far away as the shores of Greenland. The mountains of Scandinavia belong to the same Caledonian geosyncline.

The Hercynian geosyncline was bordered on the south by the High Atlas and on the north by southern England and part of Ireland. A part of the Sahara and nearly all of England and Scandinavia have now been incorporated into the platform. The Alpine geosyncline developed on an even smaller territory, encompassing only the western part of the Mediterranean Sea and the adjacent coastal mountain ranges. Many other such cases of the contraction of geosynclines during the transition from cycle to cycle can be found on the diagram of the geotectonic regions of the earth enclosed at the end of this book.

A very important feature of this process is the fact that it occurs discontinuously, with interruptions. In the course of the cycle the division of the earth's crust into geosynclines and platforms is fully maintained: the changes reflected in the expansion of the platforms at the expense of geosynclines take place intermittently at the boundaries between cycles.

This indicates that the directional development of the structure of the earth's crust and the periodic recurrences in this development are connected by certain deep bonds.

According to the majority of present-day specialists in tectonics, the

fundamental tendency in the tectonic history of the earth is an intermittent expansion of platforms at the expense of geosynclines from one cycle to another. This view, however, is not universal. Some researchers believe that geosynclines can arise in new locations that had earlier been platforms (Peyve and Sinitsyn, 1950). Nevertheless, if one keeps in mind the typical geosynclines and the entire complex of processes taking place in them, it can be affirmed that not one case has yet been established of the transformation of a platform into a geosyncline in the history of the earth. We are not, of course, speaking of the partial capturing of the margin of a platform by an expanding geosyncline (usually by the "rolling" of a foredeep upon the edge of the platform). Such cases obviously do not affect the general regularity. Some of the examples of the transformation of platforms into geosynclines, as indicated in the literature, must be considered erroneous. A. V. Peyve and V. M. Sinitsyn, for example, think that until the Alpine cycle, platform conditions existed in the Transcaucasus and in the Alps; however, the dislocations of the Paleozoic deposits that took place in the Hercynian epoch of folding, both in the Transcaucasus and in the Alps, were so great and so complex that they must undoubtedly have been produced under geosynclinal conditions. The thicknesses of the Paleozoic deposits also indicate a geosynclinal environment in both cases.

It is true that the geologic record as a rule makes it possible to determine the history of the geosyncline accurately only during the last cycle of its formation. The deposits of previous cycles (for example, of the Paleozoic in the Alpine geosynclines) are usually encountered only in isolated, unconnected districts. Thus one cannot always assert categorically that geosynclinal conditions existed in preceding cycles, and there is some room for doubt. But it must be repeated that reliable data on cases of reverse evolution still do not exist at present.

A very special type of development is observed in central Asia, in the Tien Shan region. Here typical geosynclinal conditions existed in the Precambrian, in the Early Paleozoic, and (in the southern ranges of the Tien Shan) also in the Middle and Late Paleozoic. The Hercynian, in the case of the southern ranges, and the Caledonian, in the case of the northern, are the last cycles of geosynclinal development of this region. From the end of the Paleozoic and during the whole of the Mesozoic, platform or at least parageosynclinal conditions are known to have existed, with their typically small amplitude of oscillatory movements, weak development of folding, and lack of igneous activity. From the middle of the Tertiary period, however (beginning of the Oligocene), an unexpected and very intensive "revival" took place in this region. This was reflected in a great increase in the differentiation and contrast of the vertical movements. The gentle subsidence and uplift characteristic of a platform were suddenly intensified and the upward and downward

movements in them greatly accelerated; as a result, toward the end of the Tertiary period high mountain ranges separated by deep intermontane basins rose on the site of the former platforms or parageosynclines (Petrushevskiy, 1948*a* and *b*).

The present structure of the Tien Shan could be classed with geosynclines, on the basis of the contrast in the oscillatory movements. But many of the other indications of geosynclines are not found here. It is known that a geosyncline during the first stage of its development is, in general, a region of subsidence and resulting accumulation of thick deposits, chiefly marine. The Tien Shan as a whole, with all its mountain ranges and basins, is now a region of absolute, but sharply differentiated, uplift; for this reason the sea has up to the present not been able to penetrate this area. The Upper Tertiary and Quaternary deposits, to judge by the conditions of their formation, are continental. There are still no indications in the Tien Shan of the latest folded dislocations of the continuous type, and the igneous activity is limited to very small basalt flows, which are not at all comparable to the enormous lava flows and intrusions that occur in true geosynclines. In Chap. 3 it was shown that the deeper structure of the earth's crust in the Tien Shan also differs from that in young geosynclines.

All these features argue that the Tien Shan is not a present-day geosyncline that appeared during the Tertiary on the site of a previous platform or parageosyncline. It is, apparently, rather a new form of development of the earth's crust, which is actually postplatform and resembles geosynclines in some respects but at the same time contains essential differences. Similar cases of the revival of activity on platforms by the formation of areas of intensive uplift and subsidence are encountered, though on a smaller scale, in western and eastern Siberia. Such renewed uplifts are the Altay, the western Sayan, the eastern Sayan, and a number of other ranges in the southern part of Siberia.

The laws governing such revivals of platforms are still not completely known. Regions that undergo such a renewal will be called *regions of geotectonic activation*, or *activated platform regions*, and the areas of uplift and subsidence formed in them will be called *activated*. These regions require detailed study to discover the conditions under which they arise. One can, however, apparently affirm that the activation of platforms is a new phenomenon in the history of the earth, marked by new forms of development and having no relationship to the regularities observed throughout the entire known geologic history of the replacement of geosynclinal by platform conditions.

It will be interesting to discover how later geosynclines are disposed in relation to individual zones into which earlier geosynclines had been divided. To judge from a number of examples, a regularity can be seen here in the fact that the intrageosyncline of the later cycle is located

within the limits of the subsidence in the geosyncline of the previous cycle at the very end of its development; that is, in the foredeeps and intermontane basins. In this sense there is a definite "inheritance" of new zones of subsidence from those which terminate the previous cycle. The contraction in the area of the geosynclines in the new cycle is due to the fact that not all the postinversion basins of the old geosyncline become new intrageosynclines and those that do are not "inherited" along their entire length.

As an example, one may take the transition from the Hercynian cycle to the Alpine cycle in the Caucasus (Fig. 136). In the Lesser Caucasus, Paleozoic deposits are accumulated in the central Armenian upland and

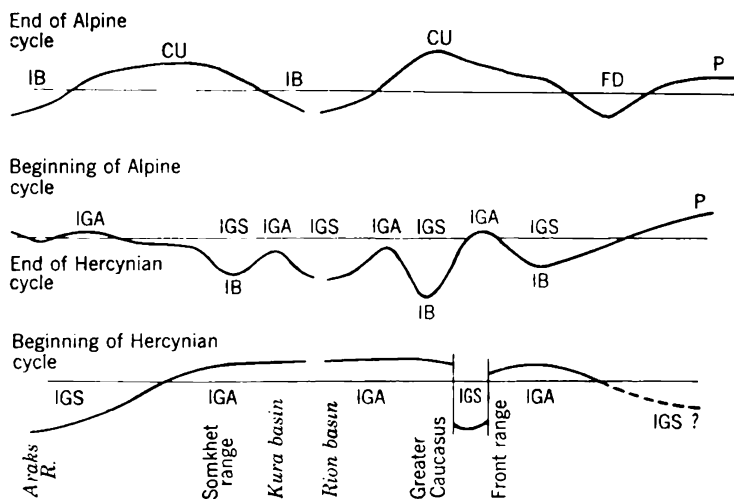


FIG. 136. Diagram of the development of oscillatory movements in the Caucasus, illustrating the transition from the Hercynian to the Alpine cycle.

totally absent in the Somkhet zone. From this, one may conclude that in the Hercynian cycle central and southern Armenia were an intrageosyncline, whereas an intrageanticline occupied the area of the Somkhet zone. At the end of the Hercynian cycle an uplift appeared on the location of the Armenian intrageosyncline; this central uplift was preserved during the Alpine cycle as an intrageanticline (the Araks IGA) in the Taurus-Caucasus geosyncline. At the same time the Somkhet zone became a marginal basin at the foot of the central uplift and was transformed into an intrageosyncline in the Alpine cycle.

In the northern Caucasus, according to D. S. Kizeval'ter, the front range during the Hercynian cycle was an intrageosyncline in which very thick Carboniferous deposits accumulated. At the end of the Paleozoic this depression was the site of an uplift, which during the Alpine cycle played the part of an intrageanticline (the Sadonian IGA)

bounded by contrasting areas of subsidence on the north and south (Fig. 136).

At the end of a cycle the basins to a considerable degree occupy the sites of the intrageanticlines of the beginning of the cycle, and such succession produces a periodic alternation of depressions and uplifts. But if at the same time certain parts of the geosyncline cease to be active, then in the transition from one cycle to another the central uplifts will be displaced toward the successively appearing active depressions. Such a migration has been observed in central Asia. According to A. V. Peyve (1947a), the Caledonian inversion took place in the northern Tien Shan (the Talasskiy Alatau) when the existing intra-geosyncline developed into a central uplift. The latter was preserved in

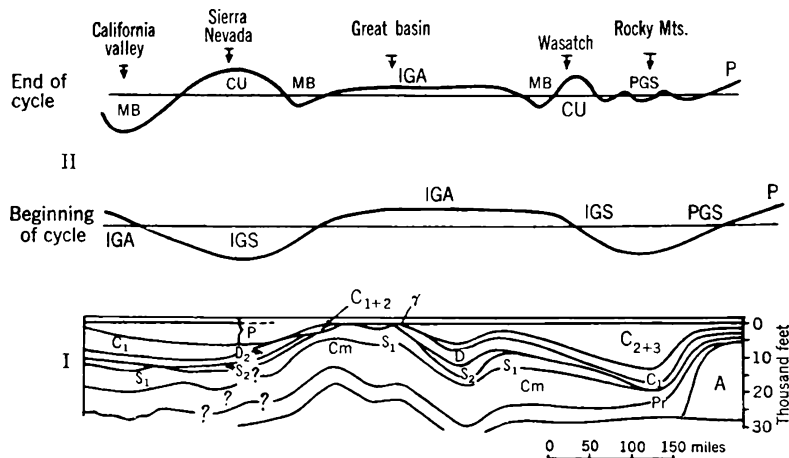


FIG. 137. Diagram of the development of oscillatory movements in the Cordilleras of North America: (I) distribution of thicknesses at the end of the Paleozoic; (II) Alpine cycle (beginning and end).

the Hercynian geosyncline as an intrageanticline, but Hercynian basins appeared north and south of it. Of these two basins, only the southern one was typical of intrageosynclines in its development: here, to the south of the Tera-Alatau deep fault, in the Ugamskiy, the Pskemskiy, and the northeastern part of Chatkal'skiy ranges, great thicknesses of Upper Silurian, Devonian, and Lower Carboniferous deposits accumulated. After the Early Carboniferous a Hercynian central uplift rose here, and an intermontane basin was formed on the site of the Fergana Valley, which had previously been an intrageanticline. The development of the Tien Shan in the Caledonian and Hercynian cycles is shown diagrammatically in Fig. 120.

An alternation of areas of uplift and subsidence in time, similar to that which was described above in the Caucasus, is observed in the Cordillera of North America. Figure 137 (I) shows a stratigraphic section through

the Paleozoic deposits of the Cordillera. From the distribution of thicknesses it appears that toward the end of the Paleozoic there were two depressions in the Cordillera filled with thick deposits of mainly Carboniferous age. Between these areas of subsidence there was an uplift in which the coal measures wedge out to a considerable degree. From the distribution of thicknesses of the Lower Paleozoic (Cambrian and Silurian), it may be concluded that the uplift resulted from an inversion in the zone of greatest subsidence at the beginning of the Paleozoic. The Hercynian central uplift became an intrageanticline in the Alpine cycle [Fig. 137 (II)], bordered on both sides by intrageosynclines inherited from basins existing at the end of the Hercynian cycle. At the end of the Alpine cycle there was a partial inversion in both intrageosynclines, and new uplifts were formed there. The central intrageanticline, however, was preserved until the end of the cycle, and only its edges were involved in the marginal basins.

This example shows a development leading to a gradual complication of the internal structure of the geosyncline. At the beginning of the Paleozoic there was a single, simple depression. By the end of the Paleozoic it was complicated by a central uplift. During the Alpine cycle new central uplifts rose in the basins remaining on both sides of the uplift. The peculiarity of this example lies in the fact that, after each inversion, the intrageanticlines did not become basins but to a considerable degree were maintained as stable zones of uplift, whose number within the geosyncline was thus gradually increasing.

This section may be concluded with one more observation: it is not impossible that the regular development of intrageosynclines and intrageanticlines of different orders may lead to somewhat different results. This has been pointed out by V. Ye. Khain (1946). It must be kept in mind that the regularities established in this chapter apply to the division of the geosyncline into its major intrageosynclines and intrageanticlines and into zones of uplift and subsidence of the first order. These zones may be complicated by secondary or smaller uplifts or subsidences, whose development may have different results. This whole problem requires further study.

DEVELOPMENT OF OSCILLATORY MOVEMENTS ON PLATFORMS

The Russian Platform

The regularities that appear in the development of oscillatory movements on platforms may be introduced through the example of the platform that has been most thoroughly studied: the Russian platform, encompassing the European part of the U.S.S.R. Platform conditions have existed here since the beginning of the Paleozoic and continue into the present time. Many investigators have studied the tectonic history of

the Russian platform. The classic works by A. P. Karpinskiy (1883, 1887, 1894, 1919) are devoted to this subject. The greatest advances in the knowledge of the tectonics of the Russian platform were made by the researches of A. D. Arkhangel'skiy (1923, 1932, 1934, 1937, 1941, 1948, etc.), and N. S. Shatskiy has devoted a number of his major works (1937, 1945*a*, 1945*b*, 1946*a*, 1946*b*, 1947, 1948, etc.) to this platform. The present author (Belousov, 1944) has made a special study of the history of oscillatory movements on the Russian platform. The investigations of A. B. Ronov (1949), who applied new methods to the study of the history of the platform, are extremely important. The latest researches into the development of the Russian platform, taking account of the most recent data, have been carried out by M. F. Mirchink and A. A. Bakirov (1951).

Before the beginning of the Paleozoic, the territory of the Russian platform, bordered on the east by the Ural geosyncline and on the northwest by the Grampian geosyncline in Scandinavia and extending southward into the area of the Azov-Podolian massif and the Ciscaucasus, was uplifted and became an extensive region of erosion. Some parts of this platform began to subside at the beginning of the Paleozoic. Until recently Lower Paleozoic deposits within the Russian platform were known only in the Baltic Sea area; now, however, as a result of deep drilling, more extensive occurrences of Lower Paleozoic deposits have been found. A zone of these deposits, which has been dislocated to a considerable degree, extends from the Baltic Sea northeastward into the basin of the northern Dvina River and, possibly, curves around to the east of the Baltic crystalline shield and disappears northward into the Barents Sea. Another Paleozoic zone is observed between Moscow and Saratov, which apparently widens along the lower reaches of the Volga River; its Lower Paleozoic deposits supposedly include a large area in the Caspian depression. Finally, a third zone of Lower Paleozoic deposits follows the western slopes of the Urals from the Pechora River basin to the source of the River Kama.

In view of the area with a small accumulation of sediments, which was apparently repeatedly interrupted, the Baltic zone stretched southward approximately as far as Orel. Further to the south is a large area in which there are no occurrences of Lower Paleozoic sediments. This region includes the Voronezh and the Ukrainian massifs, which were one unit at that time. A region of uplift corresponding to the Baltic shield bordered this zone of Paleozoic deposition on the northwest. In addition, in the central part of the Russian platform, in the Middle Volga region there is a fairly large area where Lower Paleozoic deposits are lacking. Deposits of this age are also absent on the Timan.

Everywhere except in the Baltic Sea area, the Lower Paleozoic on the Russian platform is represented by clastic beds of continental origin,

which contain no fossils. The stratigraphy of these beds has not been sufficiently studied, and the regions in which they occur are far from accurately contoured. In the vicinity of the Baltic Sea the Lower Paleozoic is represented mainly by marine deposits of varied composition. The Lower Cambrian is composed of terrigenous rocks and the

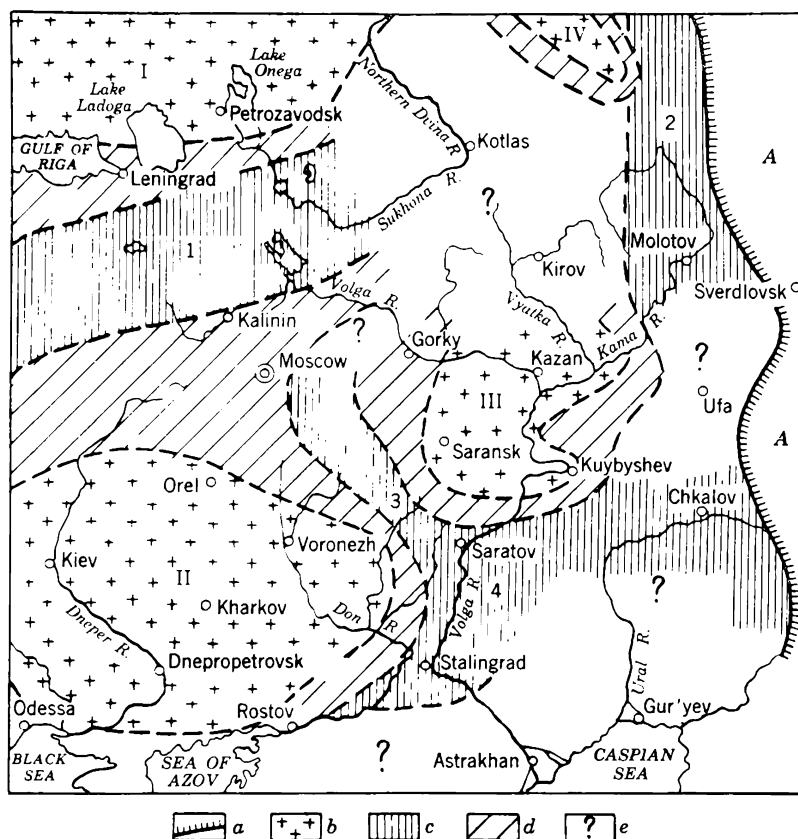


FIG. 138. Tectonic diagram of the Russian platform during the Early Paleozoic: (a) boundary of the Ural geosyncline; (b) regions of uplift (antecises); (c) regions of subsidence (synclises); (d) intermediate regions, characterized by alternation of uplift and subsidence; (e) unexplored regions. Antecises: (I) Baltic; (II) Ukrainian; (III) central Russian; (IV) Timan. Synclises: (1) Baltic, (2) Kama-Pechora, (3) Saratov-Moscow, (4) Caspian (after M. F. Mirchink and A. A. Bakirov). Original authors' terminology changed.

Lower Silurian mainly of limestones, and dolomites are predominant in the Upper Silurian. During the Middle and Late Cambrian the accumulation of sediments was interrupted.

The regions of accumulation of Lower Paleozoic deposits naturally correspond to the areas of subsidence of the earth's crust, and the places that lack these deposits are the sites of uplifts. Figure 138 shows

a diagram of the locations of subsidence and uplift on the Russian platform during the Early Paleozoic. Because of the small amount of study of these sediments in the central and eastern parts of European Russia, this diagram is to a great degree hypothetical. One may speak of the existence, during the Lower Paleozoic, of the Baltic, the Kama-Pechora, the Saratov-Moscow, and the Caspian subgeosynclines and of Baltic, Ukrainian, central Russian, and Timan subgeanticlines. No isopachs have been drawn because of insufficient data.

At the end of the cycle in the Baltic Sea area, the earth's crust was elevated in connection with the end of the cycle in the Caledonian geosyncline of Scandinavia and the formation of a mountain region in its place. The wave of uplift spread from here far to the southeast and involved the northwestern part of the Russian platform. Moreover, by the Devonian period, in the western half of the platform the zone of maximum subsidence and deposition of sediments shifted southward and its axis passed through Moscow. This was the origin of the Moscow subgeosyncline, which first appears clearly in diagrams of the distribution of thicknesses and facies of the Devonian deposits (Fig. 139). The Ukrainian subgeanticline was divided into two regions of uplift—the Voronezh and the Azov-Podolian—between which was the Donetsk depression, which later went through not a platform but a parageosynclinal development. The area of the central Russian subgeanticline decreased, and in connection with this, the eastern Russian subgeosyncline was formed. This was divided at that time by a broad lateral bend into two smaller subgeosynclines, the Caspian and the Kama-Pechora. Between these subgeosynclines and the Ural geosyncline, with its great thicknesses of Devonian rocks, there was a meridional zone of relatively small thicknesses: the Ural subgeanticline.

By the beginning of the Carboniferous period the central Russian subgeanticline had ceased almost entirely to act as a barrier, and from that time the eastern Russian subgeosyncline became a single zone of subsidence, which was only a little smaller in a lateral belt from Kuybyshev to Ufa (see Fig. 90).

This division of the platform, reflecting the positions of more or less complete geologic sections, was maintained until the end of the Carboniferous (see Fig. 91). The changes during that period consisted at first of a very great subsidence developing in the Donetsk parageosyncline. Secondly, in connection with a great uplift in the Ural geosyncline, where a mountain range was formed, a foredeep appeared in the Ural area; this migrated westward and drew into its area of subsidence the zone in which the Ural subgeanticline had been situated earlier. Other changes also took place along the boundaries of the subgeosynclinal and subgeanticlinal regions, but these were of no importance in regard to principles.

Greater changes occurred at the beginning of the Permian. During this period, in the central part of the Russian plain, there was a general and significant growth of the subgeanticlines at the expense of the Moscow and the eastern Russian subgeosynclines (see Fig. 92). Toward the end

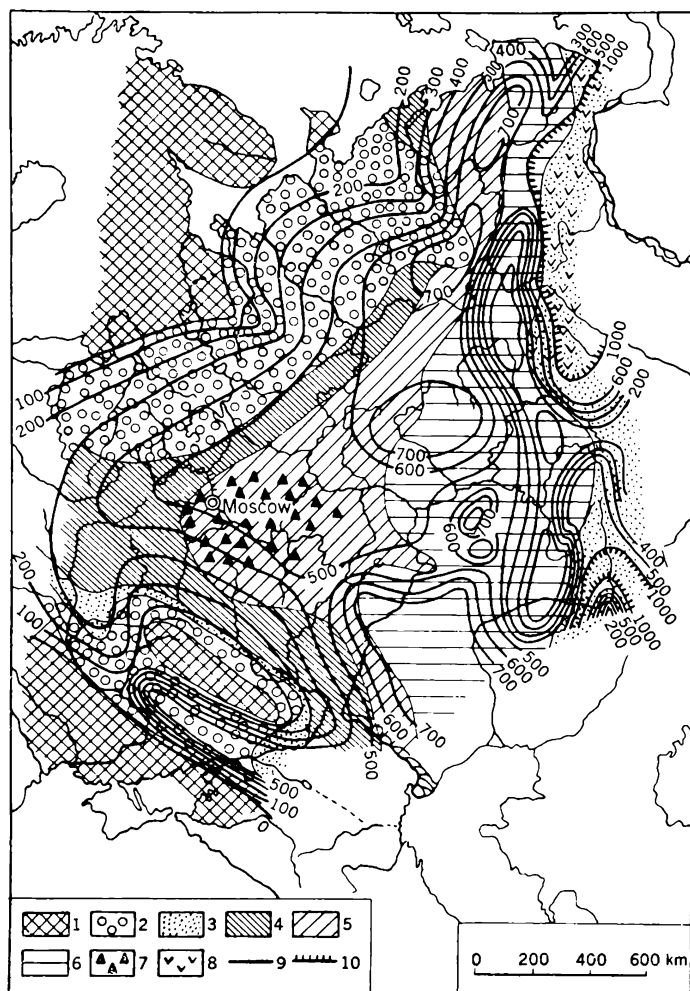


FIG. 139. Facies of Famennian formations and thicknesses of Upper Devonian deposits on the Russian platform: (1) regions of erosion; (2) red-colored deposits; (3) sands and clays; (4) clays, limestones, and dolomites; (5) limestones and dolomites with interbeds of marls and clays; (6) limestones and dolomites; (7) gypsum; (8) volcanic rocks; (9) isopachs at 100-m intervals; (10) isopachs at 1,000-m intervals (after A. B. Ronov).

of the Permian (in the Tatarian age) this process resulted in a zone of zero thicknesses spreading eastward beyond the Volga, and the accumulation of sediments in the Moscow subgeosyncline ceased almost completely (Fig. 140). By that time, too, the whole Ural geosyncline

was strongly uplifted together with its foredeep; there was also a considerable upheaval in the Donets paraegeosyncline, on the site of the folded Donbas. The part of the zone of subsidence that was preserved (north and northwest of the present Donbas) began to act as a marginal basin in relation to this uplift. Consequently, by the end of the Permian and beginning of the Triassic, subsidence and accumulation

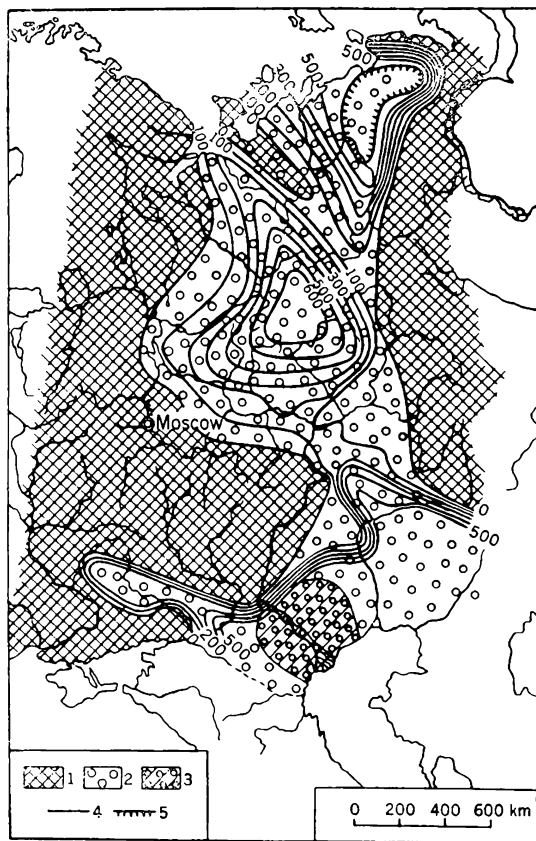


FIG. 140. Facies and thicknesses of the Tatarian stage on the Russian platform: (1) regions of erosion; (2) red-colored deposits; (3) red-colored deposits with marine interbeds; (4) isopachs at 100-m intervals; (5) isopachs at 1,000-m intervals (after A. B. Ronov).

of sediments continued only in the contracted eastern Russian subgeosyncline, in the aforementioned Donbas foredeep. Toward the Middle Triassic the continued growth of the uplifts from the east and west led to a total elevation of the whole area of the platform and its transformation into a region of erosion.

At the beginning of the Mesozoic, in the location of the Ural geosyncline, a young platform area was formed, joining together the Russian

and Siberian platforms, which had earlier been separate. The Urals, which had continued to rise gradually, now became a subgeanticline. The other geotectonic zones in the European part of the U.S.S.R. will be described in the discussion of the thicknesses of the Mesozoic and Cenozoic deposits.

At the beginning of the Jurassic, following Triassic uplift, the Russian platform again sank. This time the subsidence spread from the south—from the borders of the Caucasus geosyncline, which was subsiding intensively at that time. The subsidence, which caused a marine transgression, encompassed the Lias in the southeastern and southwestern part of the Russian plain (the Ural-Emba and the Dnepr-Donets regions). In the Dogger it spread northward along the Volga and by the beginning of Late Jurassic covered the greater part of the plain. This subsidence was also considerably less than in the geosyncline, which was now the Caucasus geosyncline.

The distribution of areas of greater or less accumulation of sediments in the subgeosynclines and subgeanticlines is shown in the diagram of the distribution of Upper Jurassic thicknesses (see Fig. 93). This shows approximately the same zones in the central part of the Russian plain that appeared in its Paleozoic history: the eastern Russian and Moscow subgeosynclines and the Voronezh and Baltic subgeanticlines. But the widening of the platform along its margins caused new zones to arise: the Ural subgeanticline already mentioned, and farther south the Voronezh subgeanticline on the site of the previous foredeep of the Hercynian Donets parageosyncline (which was not renewed in the Alpine cycle), and the Dnepr-Donets subgeosyncline, which was bordered on the south by the Azov-Podolian subgeanticline and merged in the southeast into the Ciscaucasus (Stavropol) subgeanticline.

The same disposition of subgeosynclines and subgeanticlines was maintained in the Cretaceous. The second half of the period saw a rapid growth of the subgeanticlines and their uplift to the north; in the south, the subsidence and expansion of the subgeosynclines were intensified. Thus the Dnepr-Donets subgeosyncline was greatly enlarged, forcing the Voronezh subgeanticline back northward, into the Tula region (see Fig. 94). In the Tertiary the uplifts in the north continued to grow, and the zones of subsidence were concentrated more in the south. Subsequently, these were preserved only in the foothills of the Caucasus and in the Crimea—in the foredeeps of the Alpine geosyncline. During the Quaternary there was again a general uplift of the Russian plain, and it once more became a region of erosion.

The latest movements on the Russian platform have been studied by Yu. A. Meshcheryakov (1950, 1951). His researches have established that the division into subgeosynclines and subgeanticlines was maintained in the general uplift of the platform at the end of the Alpine

cycle, taking the form of less uplift (relative subsidence) of the subgeosynclines and greater uplift of the subgeanticlines. Thus the elevation of the platform as a whole at the end of the cycle is described as a more general process than the movement of the individual subgeosynclines and subgeanticlines. This inheritance of present-day movements from older ones is especially clear on the margins of the platform; but its central areas during the Quaternary period were gradually drawn up in a broad anticlinal uplift spreading outward from the Baltic shield, which imposed a certain common character on the movements in this part of the platform and led to its subdivision, less sharply defined than before, into areas of uplift and subsidence.

From what has been set forth above one must conclude that the platform is by no means an immobile block. Within the limits of the platform, the earth's crust undergoes varied and complicated oscillatory movements, whose development on the whole shows a certain periodicity. The complex movements, beginning with the Cambrian, fall into three cycles, each of which starts with subsidence and ends with uplift. Subsidence of the earth's crust occurs at the beginning of the Cambrian, when a marine transgression developed in the Baltic Sea area and sediments began to be deposited in other regions of the platform. At the end of the Silurian there was an uplift of the earth's crust, most clearly manifested in the Baltic area. New subsidence and a new transgression developed in mid-Devonian. This cycle terminated in a general uplift of the platform at the end of the Permian, when the whole platform became dry land. The last cycle started with subsidence and a marine transgression in the Jurassic and ended with uplift and retreat of the waters from the platform at the end of the Tertiary and in the Quaternary.

These cycles correspond, respectively, to the Caledonian, Hercynian, and Alpine cycles distinguished above; they occur on the platform at the same time as the latter take place in the geosynclines. Haug (1900) formulated a "law," according to which movements in geosynclines and on platforms take place, as it were, with opposite signs: during the subsidence in the geosyncline, the platform is uplifted; and during the uplift in the geosyncline, the platform sinks. A. D. Arkhangel'skiy has shown the incorrectness of this "law"; from the evidence cited here one may also see that it does not correspond to reality.

It will be observed, however, that both the downward and upward movements on the platform are usually somewhat retarded in relation to those in the neighboring geosynclines. The geosyncline in Scandinavia apparently began its subsidence as early as the end of the Proterozoic (in the deposition of the Sparagmite series), whereas in the Baltic Sea region the beginning of subsidence is dated as Early Cambrian. A wave of uplift in the same geosyncline, connected with the inversion, appeared

at the beginning of the Silurian, but in the Baltic part of the platform the uplift began only at the end of this period. The Hercynian cycle in the Urals began with the Devonian; on the Russian platform the subsidence and deposition began at the end of the Middle Devonian and continued into the Early Triassic, long after the development of the Ural geosyncline had ended. Subsidence in the Alpine cycle, which began on the platform in Early Jurassic, gradually spread during this period over the whole territory of the Russian plain, and in the south of the plain subsidence continued at a time when the Caucasus was already being uplifted.

When there was subsidence within the platform and sediments accumulated on it, the distribution of their thicknesses and facies shows that the platform was divided into subgeosynclines and subgeanticlines. But the distribution of these has by no means been constant. The present writer cannot agree with the investigators who believe that the regions of uplift and subsidence on the platforms are "fixed" in location by the deep faults that border them (Shatskiy, 1945). The boundaries between the zones of uplift and subsidence change continually throughout geologic history.

The changes that have occurred on the Russian platform during the three cycles described bear a particular character and differ from those in geosynclines. First, the general distribution of subgeosynclines and of subgeanticlines, seen in its rough features, is more stable: it is maintained during all three cycles. As early as the Caledonian cycle, the eastern Russian subgeosyncline (consisting of the Kama-Pechora and the Caspian smaller syneclises: see Fig. 138) and the Moscow subgeosyncline appear in outline, although the latter's axis of maximum subsidence was then not in the same place as now but farther northwest, in the Baltic shore area. This arrangement—the eastern Russian and Moscow subgeosynclines, with the characteristic, gulflike contours of the latter and their bordering subgeanticlines, the Baltic and the Voronezh is seen in all succeeding schemes. This fact has already been pointed out in Chap. 13. The changes in it were not significant, and they were clearly due to the influence of the neighboring geosynclines on the platform. The uplifts taking place at the end of the Caledonian cycle in the Scandinavian geosyncline spread in a wavelike motion upon the Russian platform. With this wave of uplift from the northwest is associated the displacement of the axis of the nearby subgeosyncline to the southeast and, by the same token, the transformation of the Baltic shore into the Moscow subgeosyncline. At the end of the Hercynian cycle, a foredeep was formed at the foot of the rising Urals and migrated toward the platform. Thereafter the wave of uplift from the Urals penetrated deep into the platform and caused the narrowing of the eastern Russian subgeosyncline. At the beginning of the Alpine cycle, subsidence and

marine transgression spread northward from the nearest Alpine geosyncline, the Caucasus.

It is generally observed that in each cycle the most active zones of uplift and subsidence on the platform are those parallel to the geosyncline that is "active" in the given cycle. Thus, in the Caledonian cycle the most intensive movements are in the Baltic subgeosyncline extending from southwest to northeast parallel to the Scandinavian geosyncline; in the Hercynian cycle they are in the meridional eastern Russian subgeosyncline that parallels the Ural geosyncline. In the Alpine cycle all the basic movements were associated with the southern part of the eastern Russian subgeosyncline and with the Dnepr-Donets subgeosyncline. These two zones together made up a belt parallel to the Caucasus. This redistribution of the most active zones of uplift and subsidence on the Russian platform was brilliantly demonstrated long ago by A. P. Karpinskiy (1894).

The formation of the Donets basin in Late Devonian must also be attributed to external influences. This area of subsidence appeared in association with the formation to the southeast of a Hercynian geosyncline whose subsidence spread partly into the area of the platform. The nature of this kind of influence of the geosynclines on platforms will be discussed again below.

In the second place, there have been "interior" changes associated with the "life" of the platform itself, manifested in the displacement of the zones of maximum subsidence of the subgeosyncline, in the alternation of the boundaries between subgeosynclines and subgeanticlines, and in the fact that the relative importance of both has changed with the passage of time. Among such changes are the decrease and ultimate complete disappearance of the central Russian subgeanticline between Early Paleozoic and Carboniferous, as well as the northward displacement of the Voronezh subgeanticline (from Voronezh to Tula) during the Mesozoic. Other irregularly periodic changes have been the repeated reductions and enlargements of the areas of particular zones of uplift and subsidence and their variable lateral displacements.

The essential difference between the development of a platform and that of a geosyncline is that the process occurring on the platform is not analogous to local inversions in geosynclines: during the concluding stage of the cycle one observes not a change of subgeosynclines into subgeanticlines, but a gradual growth in the area of the already existing subgeanticlines at the expense of the subgeosynclines. The uplift of the platform during the cycle is a general elevation of the platform itself, including all the subgeosynclines and subgeanticlines without fundamental changes in their distribution or their movements. Thus we are confronted with events on two levels: the internal subdivision of the platform into regions of great or small subsidence or uplift of the

earth's crust, on the one hand, and general uplift and subsidence of the region as a whole, on the other. Of course, with the intensification of the general uplift of the whole platform, an expansion of the subgeanticline will be observed, since the particular uplift is integrated into the general uplift. But in this process the essential features of the division of the platform into subgeosynclines and subgeanticlines do not change, and they are maintained during the period of general uplift, passing into the following cycle. At present one observes the general uplift of the Russian platform that closes the Alpine cycle of oscillatory movements. It has already been indicated that the assumption of the maintenance, during the general uplift of the Russian platform, of the plan of its internal subdivision is also confirmed in the case of the present epoch. For example, the region of the lower Volga and the basin of the Nizhnyaya Pechora River are areas of smaller uplift than other regions of the platform, and thus are regions of relative subsidence. These are precisely the regions that were characterized by a subgeosynclinal regime during the earlier geologic history of the platform. The most recent times, however, have seen a decrease in the sharpness of definition of the movements begun on the Russian platform after its involvement in the anticlinal uplift that spread southeastward from the Baltic shield.

The facts above present a sufficiently complete explanation of the fundamental difference between the development of a geosyncline and that of a platform. During the cycle of geosynclinal development there is a qualitative change in the geotectonic regime, in which each intra-geosyncline becomes an uplift. But no such fundamental changes in the geotectonic regime occur on the platform: here one observes periodic movements upward and downward, and such changes in the distribution of wavelike oscillatory movements as do not violate the general plan of distribution of subgeosynclines and subgeanticlines. It will be seen later that phenomena have been observed on platforms that can be regarded as an embryonic form of inversion. Nevertheless, this does not affect the disposition of the subgeosynclines and subgeanticlines.

As in geosynclines, the primary cycles on platforms are complicated by cycles of secondary and tertiary order. On the Russian platform, for example, the full Alpine cycle, as we have seen, embraces the time from the Jurassic to the Quaternary. But after the subsidence in the Jurassic period and before the Cretaceous, a short-lived uplift occurred. A similar uplift is observed at the end of the Cretaceous period. These uplifts terminate particular subcycles: the Cimmerian in the Jurassic and the Laramide in the Cretaceous.

Within the major Caledonian cycle in the Baltic shore area, a partial Cambrian (Salair) cycle is clearly expressed. In general, the same partial subcycles are seen on platforms as in geosynclines.

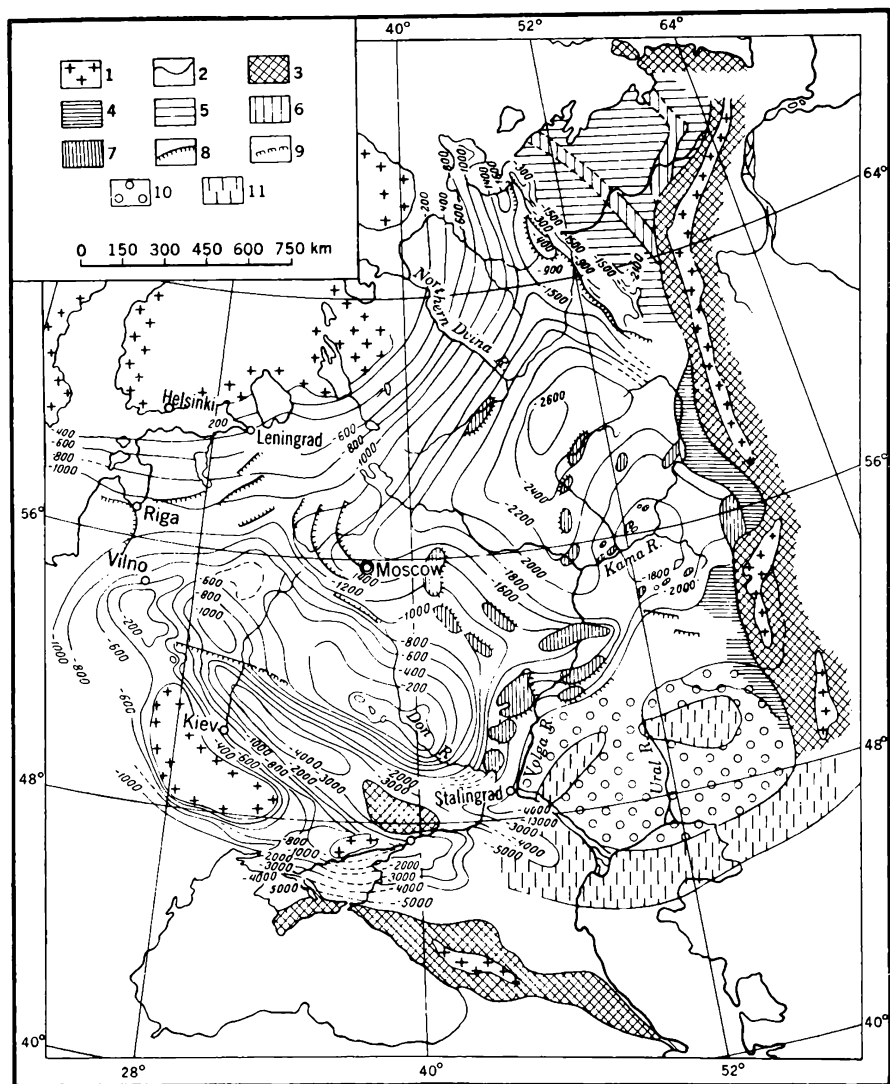


FIG. 141. Structure contour diagram of the surface of the Precambrian folded basement of the Russian platform: (1) Precambrian outcrops on the surface; (2) structure contours of the top of the Precambrian basement, in meters (from sea level); (3) Postcambrian folded structures; (4) foredeep areas; (5) Pechora depression; (6) regions of uplift in the Pechora depression; (7) Paleozoic discontinuous folds of the eastern part of the platform; (8) large structural terraces and flexures; (9) supposed fault in the northern part of the Dnepr-Donets basin; (10) salt domes of the Caspian depression; (11) areas of regional positive gravitational anomalies in the Caspian depression (after E. E. Fotiadi).

It has been indicated that the largest structural forms on platforms are synclises and anteklises; from the preceding discussion one can see that these are the structural results of oscillatory movements of the earth's crust in subgeosynclines and in subgeanticlines, respectively. Hence all the characteristic features of the structure of a synclise and an anteklise become clear. Since, for example, a synclise is the structural expression of a subgeosyncline, or zone of subsidence of the earth's crust, the thicknesses of its deposits increase regularly toward its central area of maximum subsidence and decrease toward its margins. Anteklises, formed as a result of subgeanticlinal development in which periods of absolute and relative uplift alternate, are again regularly characterized by a decrease in the thickness of the deposits toward the center until they wedge out and the anteklise becomes a region of erosion.

Since, with the passage of time, the outlines of synclises and anteklises on the map and even the wavelike alteration from one to another on the surface change, the places of maximum and minimum thicknesses of strata of different ages often do not coincide. Thus, for example, the greatest thickness of the Lower Paleozoic is in the Baltic region, whereas the Middle Paleozoic is completely absent there and its zone of maximum accumulation is in the vicinity of Moscow. Such changes cause some discrepancies in the occurrence of suites of different age on the platform. Nevertheless, because of the relatively small thicknesses and the fact that the general distribution of subgeosynclines and subgeanticlines is retained for very long periods, such discrepancies do not reach the scale that they have in the inverted structures in geosynclines. But in solving many problems of applied geology, even these relatively small deviations have to be taken into account.

Figure 141 shows a diagram of the relief on the surface of the Precambrian basement of the Russian platform, reflecting the net result of all the movements of the earth's crust since the beginning of the Paleozoic. The outlines of the main synclises (the Moscow, eastern Russian, and Dnepr-Donets basins) and anteklises (the Baltic, Voronezh, and Azov-Podolian) are clearly visible (Fotiadi, 1947).

THE INTERRELATIONSHIPS OF GEOSYNCLINES AND PLATFORMS IN SPACE AND TIME

We have seen that historically the platform structure of the earth's crust succeeds the geosynclinal structure. At the end of each cycle a certain part of the earth's crust, which had up to then been part of the geosyncline, now became part of the platform. Thus the total area occupied by platforms increased and that occupied by geosynclines decreased. Throughout the earth's surface there are a number of regions

equator but somewhat farther to the north, along the latitude of the Alpine-Himalayan mountain belt and the Caribbean Sea.

The following platforms can be identified reliably in the northern belt: (1) the Siberian, (2) the Russian, and (3) the Canadian. The southern belt contains the following platforms: (4) the Indo-Australian, (5) the African, and (6) the Brazilian.

Of the platforms enumerated, the first and fourth, the second and fifth, and the third and sixth form pairs of ancient platforms situated opposite each other. This symmetry is undoubtedly significant, but its origin is still not understood. Because of some peculiarity or other of

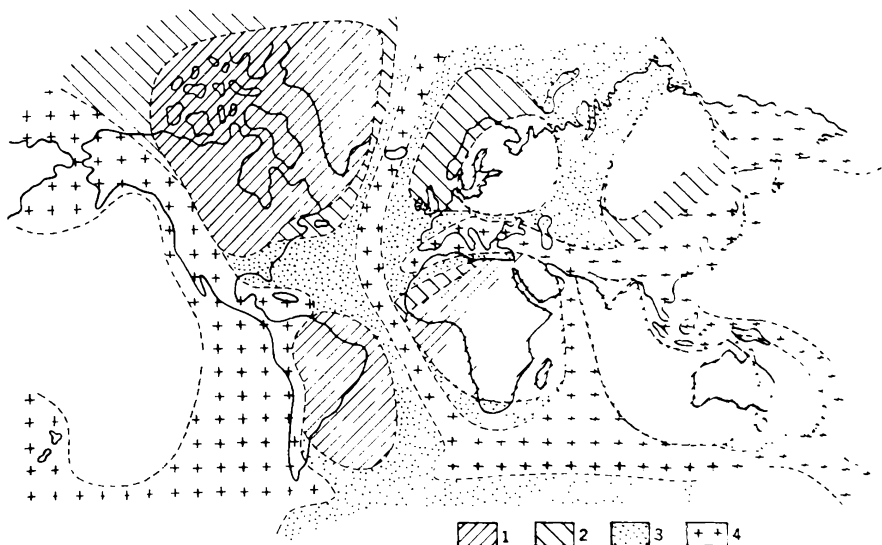


FIG. 143. Diagram of the development of platforms: (1) Caledonian platforms; (2) growth of the platforms up to the Hercynian cycle; (3) growth up to the Alpine cycle; (4) Alpine geosynclines.

the processes in the earth's interior, the transition from the geosynclinal development of the earth's crust to the platform development began in certain regularly disposed areas, from which it spread gradually to the surrounding territories.

It should, however, be noted that the transition to platform conditions occurred more rapidly in the Southern Hemisphere than in the Northern; even in Caledonian times the platforms of the Southern Hemisphere were, on the whole, larger than those of the Northern Hemisphere.

As a result of the symmetrical disposition of the ancient platforms, the geosynclines remaining outside their limits form a sort of network by the crisscrossing of their trends, which are, roughly speaking, parallel

to the earth's latitude and longitude. The Ural-Indian, the Atlantic, the Andean, and the Kamchatka-Australian geosynclinal zones approximately parallel the meridians; the central zone is parallel to the latitude, and its axis stretches from the Mediterranean Sea to the Himalayas and from there through the Pacific Ocean to Central America and farther on through the Atlantic Ocean. The northern, which is not a zone but a geosynclinal region, embraces the Arctic latitudes, and the southern embraces the Antarctic latitudes.

Although, as Fig. 143 shows, the growth of the platforms during the Hercynian and the Alpine cycles is not completely symmetrical and some regions of geosynclines end more rapidly than others, nevertheless the "network" of geosynclines as a whole is preserved to the last cycle. It is interesting to trace the way in which, as a result of the different times of the transition from geosynclines to platforms, some parts of the "network" parallel to the latitude and longitude are "intercepted," so that geosynclinal "dead ends" are formed, ending in characteristic arcs such as Gibraltar, the Antilles, the South Sandwich, and others. Sometimes it is clear that these arcs are curved toward one another, as for example the Antilles and the Gibraltar arcs, suggesting that between these two regions, in previous cycles, a direct geosynclinal connection had existed.

Inasmuch as geosynclines are the places from which the subsidence of the earth's crust at the beginning of the cycle and uplift of the crust at the end of the cycle are initiated, spreading thence into adjacent regions, geosynclines are the sources of marine transgression and regression. Consequently, the symmetry in the disposition of geosynclines and platforms is reflected in a corresponding symmetry of the displacement of shore lines and of the advance and retreat of the marine basins.

Many writers have tried to discover the reason for this regularity in the disposition of the earth's main geotectonic zones, proceeding chiefly from the shapes of the geosynclines in the Alpine cycle. They have written of regular networks of fissures predetermining the location of the geosynclines, of seams in the earth's crust formed by the action of centripetal and centrifugal forces within the earth, and so on. It is now easy to understand that all their attempts were founded on a conception that was wrong in principle. The geosynclines of later cycles are in themselves no more than the remnants of the ancient universal geosyncline, and their form is determined entirely by the shapes of the expanding platforms. Inasmuch as the latter's growth is approximately symmetrical, expanding in all directions, the geosynclines, ultimately, are retained in the form of narrow zones winding between the rounded platforms. Therefore attempts to solve the problem of the reasons for the regularity in the disposition of tectonic zones on the earth must proceed not from the shapes of the geosynclines but from the locations of the ancient

equator but somewhat farther to the north, along the latitude of the Alpine-Himalayan mountain belt and the Caribbean Sea.

The following platforms can be identified reliably in the northern belt: (1) the Siberian, (2) the Russian, and (3) the Canadian. The southern belt contains the following platforms: (4) the Indo-Australian, (5) the African, and (6) the Brazilian.

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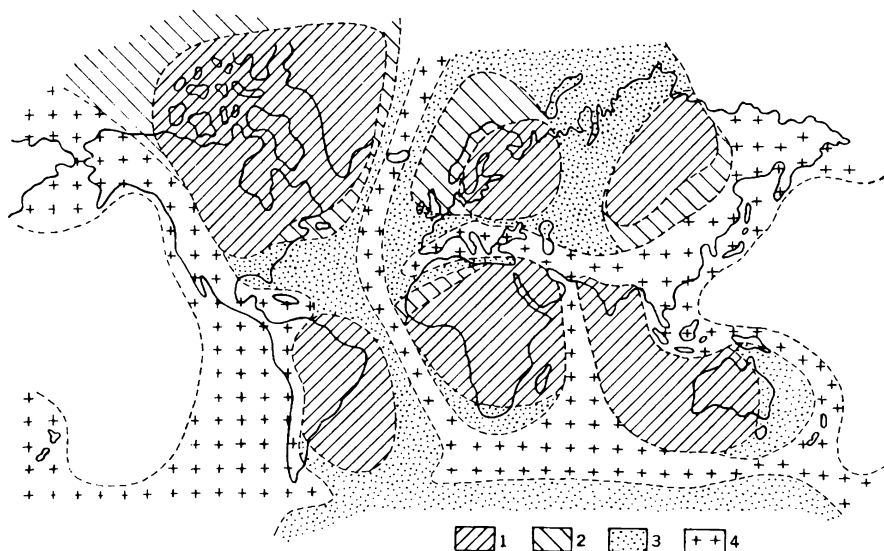


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the processes in the earth's interior, the transition from the geosynclinal development of the earth's crust to the platform development began in certain regularly disposed areas, from which it spread gradually to the surrounding territories.

It should, however, be noted that the transition to platform conditions occurred more rapidly in the Southern Hemisphere than in the Northern: even in Caledonian times the platforms of the Southern Hemisphere were, on the whole, larger than those of the Northern Hemisphere.

As a result of the symmetrical disposition of the ancient platforms, the geosynclines remaining outside their limits form a sort of network by the crisscrossing of their trends, which are, roughly speaking, parallel

to the earth's latitude and longitude. The Ural-Indian, the Atlantic, the Andean, and the Kamchatka-Australian geosynclinal zones approximately parallel the meridians; the central zone is parallel to the latitude, and its axis stretches from the Mediterranean Sea to the Himalayas and from there through the Pacific Ocean to Central America and farther on through the Atlantic Ocean. The northern, which is not a zone but a geosynclinal region, embraces the Arctic latitudes, and the southern embraces the Antarctic latitudes.

Although, as Fig. 143 shows, the growth of the platforms during the Hercynian and the Alpine cycles is not completely symmetrical and some regions of geosynclines end more rapidly than others, nevertheless the "network" of geosynclines as a whole is preserved to the last cycle. It is interesting to trace the way in which, as a result of the different times of the transition from geosynclines to platforms, some parts of the "network" parallel to the latitude and longitude are "intercepted," so that geosynclinal "dead ends" are formed, ending in characteristic arcs such as Gibraltar, the Antilles, the South Sandwich, and others. Sometimes it is clear that these arcs are curved toward one another, as for example the Antilles and the Gibraltar arcs, suggesting that between these two regions, in previous cycles, a direct geosynclinal connection had existed.

Inasmuch as geosynclines are the places from which the subsidence of the earth's crust at the beginning of the cycle and uplift of the crust at the end of the cycle are initiated, spreading thence into adjacent regions, geosynclines are the sources of marine transgression and regression. Consequently, the symmetry in the disposition of geosynclines and platforms is reflected in a corresponding symmetry of the displacement of shore lines and of the advance and retreat of the marine basins.

Many writers have tried to discover the reason for this regularity in the disposition of the earth's main geotectonic zones, proceeding chiefly from the shapes of the geosynclines in the Alpine cycle. They have written of regular networks of fissures predetermining the location of the geosynclines, of seams in the earth's crust formed by the action of centripetal and centrifugal forces within the earth, and so on. It is now easy to understand that all their attempts were founded on a conception that was wrong in principle. The geosynclines of later cycles are in themselves no more than the remnants of the ancient universal geosyncline, and their form is determined entirely by the shapes of the expanding platforms. Inasmuch as the latter's growth is approximately symmetrical, expanding in all directions, the geosynclines, ultimately, are retained in the form of narrow zones winding between the rounded platforms. Therefore attempts to solve the problem of the reasons for the regularity in the disposition of tectonic zones on the earth must proceed not from the shapes of the geosynclines but from the locations of the ancient

platforms on the earth and the process of their growth. Only after learning what conditions produced the symmetrical disposition of the initial platforms and why their development occurs regularly will one understand the origin of the shapes of the youngest geosynclines.

This description of the transformation of geosynclines into platforms is only a rough scheme that, in many of its details, is complicated by a number of subordinate conditions. First of all, a geosyncline does not always change directly into a typical platform: very often between the typically geosynclinal and typically platform stages there is an intervening stage of semiplatform or semigeosynclinal conditions, which has here been called *parageosynclinal*. Additional complications are due to the fact that not all new areas of platforms are formed immediately at the border of more ancient platforms. They also arise within geosynclines. Isolated geosynclinal regions sometimes remain within a platform. Consequently, geosynclinal, parageosynclinal, and platform regions are in places interwoven on the earth's surface in a very complicated manner: in place of the vast Hercynian geosyncline that embraced the whole of central Asia, including Mongolia, western China, part of Manchuria, and the Transbaykal, a large parageosyncline was formed during the Alpine cycle. Here the Mesozoic and Cenozoic deposits, in comparison with geosynclines, have much smaller thicknesses and are represented mainly by continental facies; here, too, the folding is mainly intermediate, and igneous activity is represented by minor intrusives. Against the background of these parageosynclinal conditions, individual parts of the area show deviations from the intermediate condition toward a greater or less intensity of development. An example of the first deviation is the eastern Transbaykal, where the large scale of the oscillatory movements and the great intensity of other tectonic and igneous processes bear witness to geosynclinal conditions. The opposite inclination—toward a more typical platform development—is found in the Ordos, in Szechwan, and many other places.

Another example of the complex interweaving of conditions, caused by the lack of uniformity in the transformation of a geosyncline into a platform, is central Kazakhstan and the adjacent parts of Siberia. Central Kazakhstan, the Kalba range, and the Altay were part of a typical geosyncline during the Caledonian cycle. During the Hercynian cycle, a large part of this territory entered into either a platform or a parageosynclinal state, but isolated regions with geosynclinal conditions, such as the Rudnyy Altay, were preserved within this territory.

The historical connection between geosynclines and platforms raises the question of the interrelationships in time and space of particular zones of uplift and subsidence within both geosynclines and platforms. Do they succeed each other, or are their interrelationships more complicated? Let us turn first to the border zones between geosynclines and

platforms. Figure 306, farther on in this book, shows a diagram of the subdivisions of part of the Alpine geosyncline of the western Alps, in which one's attention is called to the direct juxtaposition of the Briançon intrageanticline and the platform. This intrageanticline is a narrow projection of the platform extending into the geosyncline. Such direct joining of platforms to intrageanticlines is a very widespread phenomenon. In the Caucasus the Sadon intrageanticline becomes a platform along its trend in both directions. In the east this platform lies under the waters of the Caspian Sea, but in the west the relationship is very clear: the Sadon intrageanticline, which coincides with the front range, near the town of Maykop moves diagonally down the northern slopes of the Caucasus range to its foothills and then joins the platform beneath the Kuban steppes.

Such interrelations make it possible to establish a close relationship between platforms and intrageanticlines and to consider the latter to some degree as small platforms contained within the geosyncline. This view enables one to understand why the large intrageanticlines in particular are the parts of the geosyncline that, before others, change into platforms. One example is the Yukagirskoye plateau in the northeastern part of Asia, which acted as an intrageanticline in the Hercynian geosyncline and became a platform region in the Alpine cycle. This view also explains the difficulties that arise in connection with the existence of very large, undisturbed regions within a geosyncline, with the still larger intrageanticlines or even small platforms. Such are, for example, the regions of uplift and of a relatively undisturbed state of the earth's crust in the Hercynian geosyncline in central Asia. At the end of the Hercynian cycle, when mountain ranges arose out of neighboring intra-geosynclines, these regions sank and became the large intermontane basins, preserved to this time, that are so characteristic of central Asia (the Fergana, Tarim, and Dzhungarian valleys, and others). Another example is the Hungarian basin, which during the Mesozoic acted as an intrageanticline within the Alpine geosyncline. But because of its great size and its complicated internal tectonic structure it may also be considered as a small platform. The role of central Kazakhstan during the Hercynian cycle is to some extent similar.

No clear boundaries can be traced between intrageanticlines and platforms, since they apparently are genetically related tectonic elements. This relationship also appears to have a historical aspect: the "rolling" of a foredeep on the edge of a platform is completely similar to the same "rolling" of a foredeep on an intrageanticline. On the other hand, it would be incorrect to exaggerate the similarity between intrageanticlines and platforms: the intensity of the uplift of an intrageanticline, as a rule, is considerably greater than the mobility of a platform, the igneous activity observed on intrageanticlines is more violent, and

the contrasting transitions to the adjacent intrageosynclines create special "boundary conditions" for the intrageanticlines.

The geosyncline at the end of the cycle is divided into zones of uplift and subsidence. If a part of the geosyncline at the end of the cycle becomes a platform, then, as shown by observations, the previous geosynclinal zones of uplifts and subsidence, with a decrease in their intensity, will correspond in many cases to the zones of uplift and subsidence on the platform: the subgeanticlines and subgeosynclines. In these cases, one observes a direct inheritance of geosynclinal zones by particular platform zones. Such cases were formerly referred to as *posthumous movements*, but it is now more acceptable to call them *inherited*.

The widespread conception of so-called buried structures is connected with a similar inheritance. Such structures are observed where, after the replacement of geosynclinal conditions by platform conditions, the previous geosynclinal uplifts have become involved in a platform subsidence of the earth's crust. This always happens in the formation of interior depressions. In this case, under the young platform deposits of the depression, the old geosynclinal uplifts continue to be manifested in the form of a corresponding structural uplift of the folded basement, the decrease in the thickness of the platform deposits, the omission of a number of strata from the sections, and a change in facies toward greater coarseness. In other words, the platform deposits on ancient geosynclinal uplifts reveal all the characteristics of anteklises in their mode of occurrence. In reality, in such a case, the ancient geosynclinal uplift is not "buried" in the sense that some former mountain range is covered by younger sediments. The tectonic uplift that occurred in the geosynclinal stage continues to rise at a much slower rate by "inheritance" in the platform stage as well. But this rise continues as a relative uplift against the background of a greater subsidence of the interior depression. This is, for example, the relationship in the western Siberian lowland where, between the Chulym-Taz and the Irtysh-Ob' syneclises, there is a comparatively narrow linear anteklise in whose core is "buried" the Ob'-Yenisey uplift of the Hercynian geosyncline. In other cases, when the former geosynclinal zones are elevated more rapidly than the more general subsidence of the territories, the "inherited" uplifts grow absolutely and are reflected in anteklises and anticlinal uplifts subject to erosion.

The concept of the inheritance of previous geosynclinal movements by a platform greatly facilitates analysis of the structure and history of ancient geosynclines. Thanks to this, for example, it was possible to decipher to a certain degree the deep-seated structure of the western Siberian lowland and other regions (Petrushevskiy, 1951; Yanshin, 1945).

The concept of inheritance sometimes suggests interesting relationships between geosynclines and platforms. Let us imagine that geosynclinal development in the next cycle has been renewed only in some part of the intermontane basins or foredeeps of the earlier cycle and that the other parts of the basin have been joined to the platform and entered into its structure as a subgeosyncline. Then, in the following cycle, the transformation of a geosyncline into a subgeosyncline will be observed. Such cases do occur. As the distinction between a geosyncline and subgeosyncline consists primarily in the magnitude of the subsidence in the first stage of the cycle, an intermediate zone can be distinguished—a parageosyncline with an intermediate regime of movements. The geosyncline changes along its trend into a parageosyncline and then into a subgeosyncline. Such a closing off and gradual disappearance of the geosyncline along its trend may take place in both directions, thus leading to the formation of closed geosynclinal ovals, ending blindly at both ends. One such oval is the Alpine geosyncline of the eastern Transbaykal, which in both directions along its trend (both in the territory of Mongolia and in the Amur River region) changes into a parageosyncline. The Alpine geosyncline of Mexico, as it trends northward, is replaced by the parageosyncline of the Rocky Mountains, which, in turn, change into a typical subgeosyncline.

It has been said above that geosynclines, especially Alpine geosynclines, are divided into ovals. The latter, as a rule, extend along the entire length of the geosynclinal zone and may be disposed in a row, one after the other, or en echelon like the ovals of the Alpine geosyncline in Europe: the Balkan, the Hungarian, or the Mediterranean (see Fig. 113). In spite of this disposition each oval extends along the length of the axis in a direction parallel to the earth's latitude; that is, along the general trend of the whole geosyncline. Each oval terminates in a blind end along its trend, forming a curved "protrusion" into the platform. Where the geosynclinal oval is pressed along its long axis against the platform, one often finds on its continuation a zone of smaller subsidence (subgeosyncline or parageosyncline) with the same trend as the given oval and diminishing as one moves away from it. For example, the Vauconté parageosyncline, whose trend is parallel to the earth's latitude and which closes to the west and lies adjacent to the Alps, is the continuation of the Alpine oval. Another example is the Filyuy subgeosyncline in Siberia, opposite the curve of the Verkhoysk Range, which borders the enormous geosynclinal oval of northeastern Asia. In North America the outer contour of the Hercynian geosyncline borders the Gulf of Mexico and the lowlands adjacent to it in a curve (Fig. 144). This marks the edge of a great oval whose long axis runs from south-east to northwest. On the extension of this oval is a zone of subsidence that is no longer of the geosynclinal type but is parageosynclinal or

subgeosynclinal, with a thick accumulation of Paleozoic deposits: the Wichita system (Shatskiy, 1945b).

According to N. S. Shatskiy's conception, this is also the origin of the parageosynclinal depression of the Donets basin: it lies opposite an angular projection into the side of the platform of the Hercynian geosyncline, which, according to Shatskiy, lies under the Kuban lowland. N. S. Shatskiy has, in general, devoted much attention to similar depressions on the peripheries of platforms and has called them *lateral marginal depressions* (1946b). On the other hand, he explains their origin

in a different manner from the present writer, attributing primary importance to the mechanical action of stretching exerted by the geosyncline on the adjacent parts of the platform opposite the *ingoing angle* of the geosyncline. This explanation seems to be in error because the depressions in geosynclines and on platforms are equally produced by processes deep in the earth's interior, and there is no basis for imagining mechanical effects of a geosyncline on a platform or vice versa. In the present author's view, the geosynclinal oval and its parageosynclinal or subgeosynclinal continuation must be seen as a single zone of subsidence existing at the beginning of the given cycle. The central and deeper part

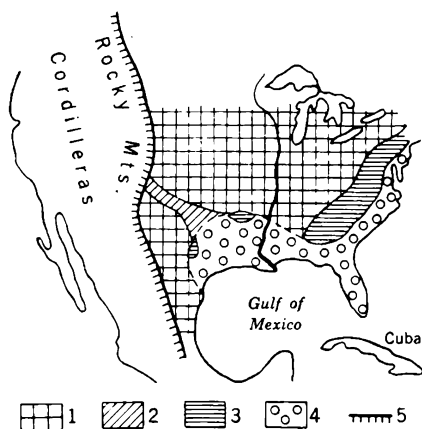


FIG. 144. Diagram of the structure of part of North America: (1) platforms; (2) Hercynian parageosynclines; (3) Hercynian marginal uplifts; (4) Hercynian interior depressions; (5) outer boundary of Alpine parageosynclines and geosynclines.

of this depression appeared to be a place in which geosynclinal conditions developed, with an inversion and the formation of a central uplift, whereas the other part, at the end of the depression where there was less subsidence, became a para- or subgeosyncline.

Nevertheless, direct inheritance of the geosynclinal areas of subsidence and uplift by the platform, as described here, is not always observed. It appears most clearly on the youngest platforms formed on the site of the geosyncline of the preceding cycle. For example, Western Europe, except for the Alps and the western Siberian lowland, consists of platforms of the Alpine cycle formed on the location of a partly Hercynian and partly Caledonian geosyncline. In the subdivision of this young platform into zones of uplift and subsidence, there are many signs of inheritance of features of the preceding geosynclinal regime. But on old platforms, which are separated from geosynclinal conditions

by one or even two cycles, such a direct inheritance is observed only locally, mainly on the borders of the platforms. This group includes the Russian, Siberian, and North American platforms, which have existed as platforms since the Caledonian cycle. Generally, in the case of the older platforms and especially their interior parts, the interrelationship with geosynclines becomes somewhat different. These relationships have been studied by N. S. Shatskiy (1948). On old platforms the regions of uplift and subsidence—the subgeanticlines and subgeosynclines—have shapes and dimensions different from those of the uplifts and depressions in geosynclines. On such platforms these regions have mainly round or irregular contours, and their diameters are hundreds and even thousands of kilometers. These features distinguish them sharply from the narrow, long, and linear zones of uplift and subsidence in geosynclines. One may contrast the Moscow and eastern Russian subgeosynclines and the Baltic coast subgeanticline, on the one hand, with the uplifts and basins of the Caucasus, on the other.

The most important fact is that these large platform regions of less intense uplift and subsidence do not merely replace the narrow and highly contrasting zones within a geosyncline as a result of a gradual leveling out and “spreading” of the latter. The subgeosynclines and subgeanticlines of old platforms represent totally different tectonic elements not related to the elements of geosynclines. They do not close within the limits of the platform but may be traced into the neighboring geosynclines, where they “interfere” with the true geosynclinal zones and can be recognized within them.

N. S. Shatskiy, the first to note this last phenomenon, pointed out two examples. The first is the southern part of the Russian platform and its connection with the Caucasus. Here certain zones of uplift and subsidence are traced within the adjoining part of the Alpine geosyncline, taking the form of lateral bends in the latter. Thus the eastern Russian subgeosyncline (its Caspian part) is traced southward into the Caspian region and further, into the territory of the Caucasus. Its presence in the Greater Caucasus is indicated by the greater thickness of Jurassic deposits and by the absence of any outcrops of rocks older than Jurassic in the eastern part of the Caucasus, as distinguished from the western. To the west of this depression, which is common to the platform and the geosyncline, there is an adjoining platform zone of uplift, whose trend is also north-south. South of this is the Stavropol uplift, which has the character of a subgeosyncline but which also extends into the territory of the Greater Caucasus, producing its main transverse anticlinal bend on the Mineral'nyye Vody meridian. Farther to the west, the region in which the floor of the Sea of Azov descends below the surface corresponds to the northwestern descent of the Greater Caucasus.

A second example is the interrelationship between the Russian platform and the Urals. East of the Russian platform is a region of uplift to which the Ufa plateau belongs. This region is not limited to the platform but has an eastward extension into the Urals, where it determines the geanticlinal discontinuity in the Ural geosyncline.

Consideration of these examples creates the impression that the division of old platforms into subgeosynclines and subgeanticlines is caused by other, more deep-seated processes than those which determine the division of geosynclines. These deep-seated processes also act on geosynclines, but in the latter case they are reinforced by very active, more surficial processes that produce the wavelike movements of the sharp, properly geosynclinal uplift and subsidence, which are often superimposed in a completely different direction and in another form on the deeper and gentler movements. The latter merely "show through" in geosynclines. But when, with the transformation of the geosyncline into a platform, the shallower movements cease, a picture of the more deep-seated movements is revealed. It will be seen later that these views can be given a foundation in theory and may become the basis for developing a hypothesis of the nature of deep-seated geotectonic processes.

From the examples of inheritance examined above, one can see that the cessation of shallow movements, with the change of a geosyncline into a platform, is not accomplished all at once. Active geosynclinal upward and downward movements die out quickly but progressively, and for a certain period they continue their "posthumous" development. Ultimately, it appears, the small uplifts and subsidences are generalized into larger uplifts and depressions typical of a platform.

There is, however, another path from the greatly contrasting geosynclinal subdivisions to the less well-defined subdivisions of a platform—the path of accelerated development, leading to a partial appearance, even on the youngest platforms, of large subgeosynclines surpassing considerably in size the zones within geosynclines. This happens in the formation of interior depressions, which often terminates the geosynclinal development of the region. Besides the western Siberian lowland, which was discussed above, an example of an interior depression is the Gulf Coast lowland adjacent to the shores of the Gulf of Mexico. This was formed at the end of the Paleozoic on the site of the Hercynian geosyncline of the Appalachians. Young interior depressions, formed on the sites of Alpine geosynclines that have finished their development, are the western part of the Mediterranean Sea, the Caribbean Sea, the interior sea of east and southeast Asia, and others. A number of examples will become familiar later on.

Thus the transition in time from a geosyncline to a platform takes many different forms. In some cases, there is an inheritance and a slow

transformation of the geosynclinal state of the earth's crust into a platform state, with a long-lasting continuation on the platform of traces of past geosynclinal conditions. In other cases, there is a rapid change to another regime of movements through the formation of large interior depressions.

In conclusion, we may touch upon some of the structural peculiarities of the areas joining platforms and folded zones. N. S. Shatskiy (1945*b*, 1946*b,c*, 1947) supposed that folded zones and platforms were separated from one another by "deep-seated fractures" and were therefore completely isolated in space. The role of deep-seated faults fixing the geotectonic boundaries in space is contradicted in this case, it seems, by the displacement of the foredeeps toward the platforms during the course of the cycle: this would be impossible if the foredeeps were "fixed" in location by fractures in the crust. The rounded boundaries between folded zones and platforms to be observed in many regions also testify against such fractures.

In most cases, the folded zone is connected to the platform by a foredeep that has the structure of a synclinorium. This foredeep, as a rule, is not symmetrical: the axis of its greatest subsidence and greatest thicknesses is usually shifted toward the folded zone. Consequently, its flank adjacent to the folded zone is steep, whereas the flank rising toward the platform is gently sloping. Both flanks may be complicated by flexures and steep steplike breaks, but these dislocations are of secondary importance.

The width of the foredeeps is different in different regions, depending on the nature of the movements on the adjacent area of the platform. The foredeeps are wider next to synclises and narrower where they are opposite anteklises. The foredeep, as shown by N. S. Shatskiy (1947), can be absent where an especially high anteklise is situated on the margin of the platform. This has been observed in Scandinavia, in the northern Appalachians, and elsewhere.

The depth of the foredeep, the amplitude of its subsidence, and consequently the thickness of the deposits accumulated in it change uniformly along its trend. Widening and narrowing, deepening and becoming shallower, the foredeep always breaks up longitudinally into a chain of "basins" separated by less deeply sunk partitions. This structure of the foredeeps was established by M. V. Muratov (1949) in the case of the Alpine foredeep in the south of the European part of the U.S.S.R. and in the Ciscarpathian region.

CHAPTER 17

FUNDAMENTAL CONCEPTS OF SEDIMENTARY SEQUENCES

It has been shown that oscillatory movements of the earth's crust proceed in a regular manner, with a definite periodicity. They result in periodic changes in the relief and dimensions of land masses and marine basins, as well as in the relationship between upward- and downward-moving parts of the crust. We also know that changes in this relationship bring about changes in the conditions under which the facies of deposits are formed. Since in any given cycle the oscillatory movements pass through approximately the same stages of evolution, it may be assumed that in each cycle there will be a more or less exact repetition of the succession of original conditions under which sedimentary strata are formed and that periodic movements will therefore result in the periodic recurrence of similar facies. Each facies should thus correspond to a specific combination of tectonic conditions characterizing a particular stage in the geotectonic cycle.

The oscillatory movements are a complex process that may be broken down into movements of different orders superimposed on each other. The overall course of oscillatory movements during a cycle is thus only a fundamental background, or average course, of the process, from which there are constant deviations to one side or the other. These secondary deviations also affect the formation of sedimentary strata. It would thus be more precise to say that each stage of a geotectonic cycle corresponds not simply to a certain facies of sediments but to a given *assemblage*, or complex of facies, that reflects both the average distribution of uplifts and subsidence and the algebraic sum of the minor oscillations.

The complex of facies in the deposits that corresponds to a given stage of the geotectonic cycle is called a *sedimentary sequence*. One may assume, consequently, that in each cycle similar sequences will be repeated during the individual stages. But the conditions under which sedimentary beds are formed, though for the most part reflecting the

relationship between uplift and subsidence, also change with the geotectonic zones. Thus it is more precise to say that each sedimentary sequence corresponds to a given stage of the geotectonic cycle and to a given geotectonic zone. There is some difference, for example, between the sequences of geosynclines and platforms belonging to the same stage of a cycle, although, of course, they also show a definite similarity.

Although it has long been known that there is a certain regularity in the distribution of sedimentary sequences in the earth's crust, the theory has actually been developed only recently in the U.S.S.R. Among the authors who have worked on problems relating to the laws of the accumulation and distribution of sedimentary sequences are Pustovalov (1940), Strakhov (1946, 1948, 1949), Shatskiy (1938, 1945a), Khain (1950b), Ronov (1949), Peyve (1948), Keller (1949), Krashennikov (1957), Rukhin (1953), Ronov and Khain (1957), the present author, and others. In spite of a number of significant achievements in this field, however, the entire problem is still in an early stage of development, and the conclusions that can now be made are still only general and approximate. Such conclusions will have to be based on the available data from observations. It has already been pointed out that the increase in subsidence relative to uplift during the first half of the cycle results in marine transgression, or an increase in the size of marine basins. The reverse process, an increase in uplift relative to subsidence in the second half of the cycle, leads to expansion of the land areas and reduction in the marine basins. Some writers distinguish *thalassocratic* periods in the earth's history, in which the sea has covered most of the earth, in contrast to *geocratic* periods, in which continental areas were predominant. A. B. Ronov has computed the areas occupied by the sea in the European part of the U.S.S.R. during various geologic periods in the last two geotectonic cycles (1949). His results, shown in Fig. 145, clearly show the periodic character of the changes in the area of the zones of deposition. A similar illustration, showing the average conditions for all continental areas during the last three cycles, appeared earlier in the book (see Fig. 111). This indicates the effect of secondary cycles, but it is qualitative in character and shows only the general trend of the process.

A considerable predominance of marine areas is characteristic of the Middle and Late Cambrian and Early Silurian, the Late Devonian and Early Carboniferous, and the Late Cretaceous. Each of these intervals of time corresponds to the end of the first half of the Caledonian, Hercynian, and Alpine cycles, respectively. The end of the Silurian and beginning of the Devonian, Permian, and Triassic periods and the end of the Tertiary and Quaternary periods are characterized by the opposite conditions, namely, the existence of large land areas and a re-

duction of the area of marine basins. These epochs correspond to the end of one cycle and the beginning of the next.

Expansion and reduction of the land areas mean expansion and reduction of surfaces exposed to erosion, and thus of the amount of terrigenous material transported to the marine basins. Thus there is a regular progression from predominantly clastic deposits at the beginning to predominantly limestone deposits in the middle and again to predominantly clastic rocks at the end of each cycle. The epochs of great marine transgressions are those during which limestone deposits reached their maximum extent, and times of marine regression were times of widespread distribution of clastic rocks, including continental deposits

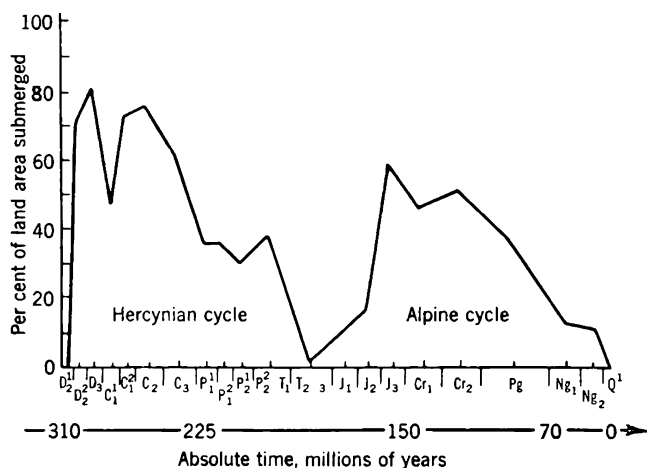


FIG. 145. Changes in the area of zones of deposition on the Russian platform during the Hercynian and Alpine cycles (*after A. B. Ronov*).

as well as deposits of lagoonal origin. These quantitative changes are shown for the Russian platform in Fig. 146.

It has been pointed out repeatedly that changes in the appearance of the sedimentary beds during a cycle depend to a large extent on climatic changes. For example, sediments such as the coal measures of the Carboniferous, the dark Silurian shales rich in organic matter, and the widely distributed limestones of the Upper Cambrian, Upper Devonian, Upper Cretaceous, and Paleogene bear clear witness to a warm, humid climate. The deposits of these series and the minerals they contain are quite uniform wherever they occur. In any case, there is no evidence of any significant change in climatic conditions on the earth's surface during the epochs in which they were formed.

On the other hand, the deposits of such times as the very end of the Precambrian, the Early Devonian, the Late Carboniferous and Early Permian, and the Quaternary testify both to a substantial cooling, even

though only of local extent, and to noticeable climatic changes. The first aspect is evident in the existence of glacial deposits and the second in the presence of warm-sea and desert sediments of the same age as the glacial deposits. It follows that the climatic changes on the earth's surface show the same periodicity as do the crustal movements. This is not surprising, inasmuch as changes in the surface relief and in the distribution of land and water are basic factors determining the climate.

At the border line between two cycles, the relief of the earth's surface displays a sharpness of form accompanied by great contrasts and by

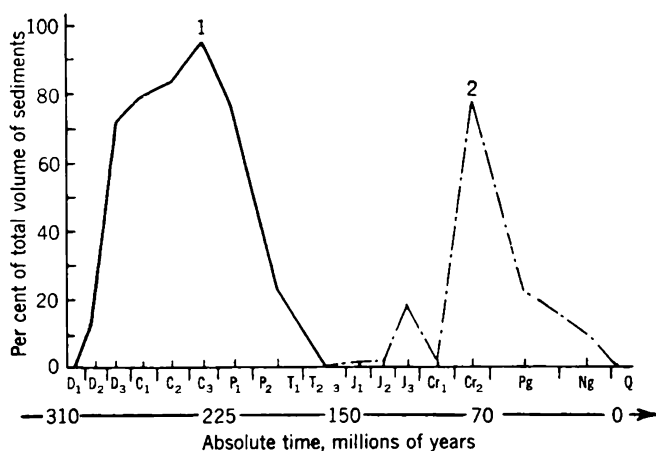


FIG. 146. Changes in the percentage of carbonate deposits during the Hercynian (1) and Alpine (2) cycles of sedimentary accumulation on the Russian platform (after A. B. Ronov).

expansion of the continents. At the end of the preceding cycle, the continents underwent great expansion and high mountain ranges appeared. This, in turn, brought about a predominantly continental climate with a wide range of temperatures and sharply differentiated hot and cold climatic zones. There is evidence both of heavy glaciation and, in the interior basins of the continents, of the existence of arid desert climates.

During the first stage of the following cycle, the uplift ceases, the land is eroded to low relief, and the continents diminish in size. This leads to the development of a humid maritime climate with small variations in temperature. During the second stage of the cycle, the uplift resumes, the relief becomes more complex, and the climate again tends to become continental.

Thus it appears that the most important factor, over geologic time, is the tectonic process, for tectonics determines climate and thus affects the character of the sediments.

The process of deposition of sedimentary sequences may now be analyzed in greater detail, starting with geosynclines.

SEDIMENTARY SEQUENCES OF GEOSYNCLINES

A comparison of the observed succession of sedimentary beds in various geosynclines, both during one cycle and during several cycles, will show immediately that there are certain regularities and that the assumption that one can distinguish the sequences corresponding to the given stages of a cycle is fully justified.

The background of the evolution of sedimentary sequences during a given cycle of geosynclinal development is provided by the previously discussed succession: from continental and terrigenous marine deposits at the start of the cycle to marine limestones in the middle and again to terrigenous and continental sediments at the end of the cycle.

The following sedimentary sequences, succeeding one another in time, may be distinguished:

1. A *lower terrigenous sequence*, which corresponds to the start of the cycle of geosynclinal development. In the Caledonian cycle it occurs in the Cambrian, especially in its early and middle epochs. In the Hercynian cycle, this sequence appears in the Early and Middle Devonian, and in the Alpine cycle, it is observed mainly in the Triassic and Early and Middle Jurassic and extends to the Early Cretaceous. This sequence is complex. At its base one often encounters continental sediments, as in the Donets basin and in the Urals (Hercynian cycle). In some cases the sequence contains thick, uniform clay or sand and clay deposits, which may overlie the continental deposits or, if these are absent, may occur at the base of the sequence. These clay or sand and clay sediments are poor in fossils but rich in organic material, either disseminated, as in oil shales, or concentrated, as in coal-bearing strata. Pyrite may also be present. The rock composition indicates a reducing environment and probably the formation of hydrogen sulfide in the basin. Some sediments of this type display cyclic bedding. One example, in the Caucasus, is the thick series of unusually uniform dark-colored shales and sandstones, containing pyrite and coal-bearing beds and dating from the Early and Middle Jurassic.

V. Ye. Khain (1950b) and B. M. Keller (1949) have called this sequence *slaty*, because of the presence of slates in it. But the "slaty" character of the slates is a result of their metamorphism, and this secondary feature should hardly be used in coining a term. A. V. Peyve apparently calls the same sequence "a sequence of terrigenous gray-colored marine deposits" (1948).

2. A *limestone sequence*, which corresponds to the end of the first half of the cycle, which is the time of the greatest dominance of subsid-

ence over uplift. This occurred chiefly in the Late Cambrian or Early Silurian during the Caledonian cycle, in the Late Devonian and the beginning of the Carboniferous during the Hercynian cycle, and in the Late Cretaceous during the Alpine cycle.

This sequence varies greatly in appearance and in its lithologic composition in different parts of the geosyncline, and even more in different geosynclines. In the Greater Caucasus, for example, the sequence includes organogenic clastic limestones, reef limestones, pelitomorphie (clay-sized) limestones, chalks, and various marls of the Upper Jurassic and Cretaceous. It also includes the so-called carbonate flysch, cyclically bedded marl, and limestone beds of the Upper Jurassic and Cretaceous that occur on the southern slopes of the Greater Caucasus.

The varying appearance of this sequence in different places makes it possible to distinguish local varieties of the limestone sequence, or subsequences, especially since the development of the latter is closely linked to tectonic conditions. The carbonate flysch, for example, is developed at the point of contact between an intensively rising *cordillera*, a narrow intrageanticline, and an even more intensively subsiding main Caucasus intrageosyncline. Other subsequences are related to less violent tectonic conditions.

The deposition of the limestone sequence preceded the inversion of geotectonic conditions and thus completed the first stage of the cycle.

3. An *upper terrigenous sequence*, which corresponds to the beginning of the second stage of the geotectonic cycle and is closely linked genetically to renewed central uplifts within the geosyncline. The accumulation of this sequence is a result of erosion of the rapidly rising uplifts.

Like the earlier limestones, the upper terrigenous sequence displays a wide variety of rocks that can be combined into several typical subsequences. One of these is characterized by a terrigenous flysch consisting of cyclically bedded sand and clay beds with occasional clastic limestone and marl strata. Examples of such terrigenous flysch series are, in the Caledonian cycle, certain Silurian graywacke beds; in the Hercynian cycle, the Upper Carboniferous and Artinskian (for the town of Artinsk in the Urals) clastic beds on the western slopes of the southern Urals; and in the Alpine cycle, the Eocene flysch that occurs widely in the Carpathians and the Alps. In all cases, these rocks accumulated at the foot of central uplifts rising within the geosynclines.

A second subsequence contains widespread mineral- and fuel-bearing beds. It consists of shales and sandstones rich in organic matter, in the form of either coal or disseminated bitumens. This subsequence reached its greatest development in the Hercynian cycle, where it is represented by the thick coal- and oil-bearing beds of the Middle and Upper Carboniferous. These are found, for example, in the Donets basin, in

the Ruhr, and in Pennsylvania. In the Caucasus geosyncline of the Alpine cycle, this subsequence is represented by the Maykopian (for the town of Maykop) bitumen-rich shale beds. In the Caledonian geosynclines this subsequence includes widespread graptolitic shales.

The two subsequences are found either together or separately (when one of them wedges out completely or almost completely). When they occur together, their relative stratigraphic position may vary. In the Donets basin the terrigenous flysch is found in the Upper Viséan stage, which is overlain by coal-bearing strata of the Middle and Upper Carboniferous. In the northern Urals, on the other hand, a coal-bearing stratum of the Upper Tournaisian and Lower Viséan lies below the Upper Carboniferous and Lower Permian flysch. The mineral- and fuel-bearing subsequence occurs in the Caucasus without terrigenous flysch.

4. The *lagoonal sequence*, characterized by a thick accumulation of common rock salt and potash, as well as gypsum, anhydrite, and dolomite. All these rocks of chemical origin were precipitated under lagoonal conditions through the evaporation of sea water.

Lagoonal deposits are widespread. In the Caledonian cycle they include the Silurian gypsum and dolomites of the Baltic region. In the Hercynian cycle, the Permian salt and gypsum beds, such as the salt and gypsum bed of the Kungurian stage in the western foothills of the Urals, are typical lagoonal sequences. In the Alpine cycle, salts and gypsum are associated with the Tertiary deposits of the Carpathian foothills and many Alpine mountain ranges.

There are, however, geosynclinal sections in which the lagoonal sequence is absent, thus producing a direct juxtaposition of the upper terrigenous formation (bottom) and the molasse formation (top). This occurs in the northern Caucasus, where the lagoonal formation is absent in the Tertiary and the mineral- and fuel-bearing subsequence of the Maykopian deposits is overlain directly by Miocene molasse deposits. On the other hand, a salt-bearing sequence does occur in the valley of the Araks River in the Transcaucasus. Such variations can be fully explained in terms of local climatic conditions.

5. The *molasse sequence*, which completes the succession of geosynclinal sedimentary sequences. It corresponds to the final stage of geosynclinal development, in which high mountain ranges are formed and uplift greatly predominates over subsidence. Terrigenous sediments predominate in the molasse formation, and the limestones are negligible. Clay material predominates in the lower strata. As one moves upward in the section, the rocks gradually become coarser until sandstones predominate. The topmost beds of the molasse sequence often consist of very coarse rocks (coarse sands and conglomerates). As a general rule, the lower part of the molasse sequence is marine and the

upper continental in origin. Molasse sediments are typically gray-colored and locally reddish. The formation contains accumulations of brown coal.

There are many examples of the molasse sequence. In the Caledonian cycle, it includes the Old Red Sandstone of the Upper Silurian and Lower Devonian, which occurs on the periphery of nearly all Caledonian geosynclines. In the Hercynian cycle, the molasse sequence is represented by the New Red Sandstone and various other red-colored Permian deposits in the western foothills of the Urals, the Donets basin, the Caucasus, Western Europe, and elsewhere. In the Alpine cycle, the sequence includes terrigenous beds of the Neogene, partly marine but largely continental, which adjoin the mountain ranges of the Alpine system: for example, the Siwalik formation in India and the adyrs (foothill ridges) of the Fergana Valley.

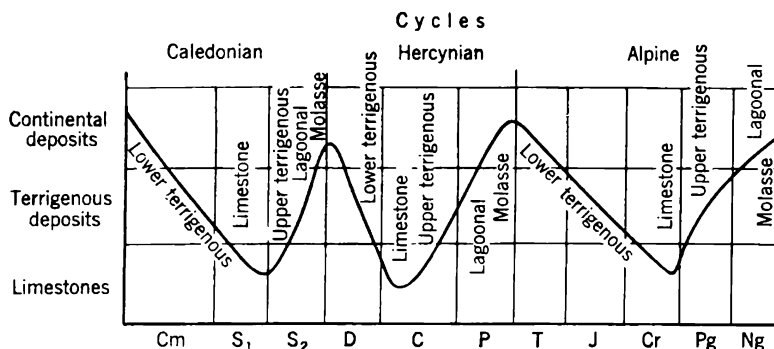


FIG. 147. Succession of sedimentary sequences during the Caledonian, Hercynian, and Alpine cycles.

6. The formation of superimposed troughs and interior basins stands by itself and, strictly speaking, does not belong among the geosynclinal sequences. It is the first of a series of platform sequences, but it will be discussed here because the formation of interior basins has been regarded as the final step in the development of the geosyncline.

Superimposed troughs generally contain continental sand and clay deposits. Basins with large reserves of brown coal are typical. Examples are the Chelyabinsk brown coal basins and basins in the Carboniferous superimposed troughs of Western Europe.

Large interior basins contain not only continental but also marine deposits, predominantly clay and sand of shallow-water origin. There are frequent occurrences of brown coals, petroleum, and natural gas. The Gulf Coast of the United States is an example of an oil-bearing region forming part of an interior basin.

A diagram of the distribution of geosynclinal sedimentary sequences in time appears in Fig. 147.

It should be explained why this survey of geosynclinal formations did not assign a separate position to the flysch, a rock type that is properly regarded as typical of geosynclines. The omission is due to the fact that this classification has tried as far as possible to observe the succession of facies and to assign each sequence to a given stage of geosynclinal development. The flysch has no definite place in such a classification because it is associated with various stages of geosynclinal development. Flysch deposits occur in the lower terrigenous sequence (at the beginning of the geosynclinal development), in the limestone sequence (toward the end of the first half of the cycle), and, finally, in the upper terrigenous sequence, which comes after the inversion.

This wide range in the occurrence of flysch deposits in the section through the deposits of geosynclines shows that the flysch corresponds

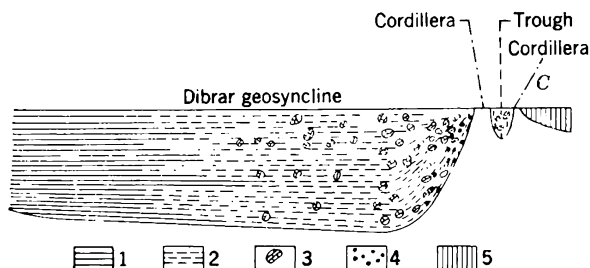


FIG. 148. Diagram showing the accumulation of flysch under conditions of greatly contrasting oscillatory crustal movements. The Dibrar downwarp of the southeastern Caucasus in the Barremian age: (1) flysch; (2) clays; (3) Upper Jurassic limestone boulders slipping down from the cordillera; (4) boulder conglomerates; (5) limestone and marl sediments (from V. A. Grossgeym).

not so much to a given stage of geosynclinal development as to more general tectonic conditions that may be characteristic of almost any stage of development. These tectonic conditions have been fully studied by N. B. Vassoyevich (1946, 1951), who showed that the formation of a flysch series reflects a contrasting combination of adjoining zones of intensive uplift and erosion, on the one hand, and equally intensive subsidence, on the other (Fig. 148). The zone of rapid subsidence next to the rising cordillera shows a rhythmic sequence of flysch beds, reflecting constant and periodic changes in the relationship between the rate of uplift and the rate of subsidence.

It is known, however, that contrasting combinations of uplift and subsidence are, in general, typical of any stage of geosynclinal development. Hence flysch deposits may be found at various levels of the geosynclinal section. This conclusion is supported, moreover, by the fact that flysch-type deposits, in the sense of a cyclic alternation of a

given limited assemblage of rocks in geosynclines, are not limited to beds classified as typical flysch. The cyclic succession of deposits also typically occurs in the molasse formation. Such flyschlike deposits of the molasse formation have been called *schlieren* by some non-Soviet geologists.

The intermittent occurrence of flysch-type deposits is thus typical of the geosyncline as a whole but not of any given stage of its development, although all terrigenous flysch is now usually assigned to the stage immediately following the inversion.

Another general indication of various geosynclinal sequences is the wide occurrence of traces of ancient buried submarine landslides. Traces of such slides take various forms: the chaotic piling up of rock debris, the presence of breccias, unusually complex and irregular "flow" deformations of strata, and so forth. These disturbed zones differ from tectonic dislocations in their restricted local character and the absence of regular geometric forms in the deposits. Some of these disturbed zones, however, may be confused with disharmonies of tectonic origin. Buried landslides are found in all geosynclinal sequences, predominantly in clay and marl sediments. They are evidence of the above-mentioned contrasting movements and, at the same time, of a dissected relief.

It was established earlier that the process of geosynclinal development is associated with a displacement of the zones of uplift and subsidence. As a result, sequences corresponding to various stages of the cycle accumulate in various parts of the geosyncline. Sediments of different ages accumulate in the geosyncline in the form of lenslike bodies that are displaced relative to each other. Therefore each sequence or group of sequences forms a lenslike body with maximal thickness where the deepest subsidence occurred at the time of deposition. It is therefore possible to distinguish the following groups of sequences: (1) pre-inversion sequences, which reach their maximum thickness in intra-geosynclines; (2) sequences of marginal deeps, which accumulate right after the inversion in zones of subsidence at the foot of the central uplift; (3) sequences of intermontane basins and foredeeps, which reach their maximum accumulation where the preinversion sequences have their minimal thickness (in intrageanticlines, IGA); (4) sequences of interior basins and superimposed trough.

It is easily seen that sequences 1 (lower terrigenous) and 2 (limestone) belong to group 1, sequence 3 (upper terrigenous) to group 2, and sequences 4 (lagoon) and 5 (molasse) to group 3. The position of group 4 is evident from its name.

The distribution of sequences in the geosynclinal section at the end of the cycle is shown diagrammatically in Fig. 149. The figure illustrates a case where the geosyncline at the end of its development contains the

following zones: foredeep; marginal uplift; interior basin; marginal uplift; intermontane basin; central uplift; foredeep.¹

Here the preinversion sequences are developed within preinversion intrageosynclines; that is, in the present interior basin and the central uplift. In both locations they have been heavily eroded as a result of uplift that continued until the end of the cycle in the central uplift and occurred right after the inversion in the interior basin. The thickness of the preinversion beds decreases and the coarseness of the facies increases toward the marginal uplifts and toward the intermontane basins and foredeeps. In the interior basin, the eroded surface of the preinversion sequences is overlain by an interior basin sequence. Thus there is a large gap in the section. It shows no deposition corresponding in time to the foredeep sequences, because at that time a central uplift occupied the place of the present interior basin.

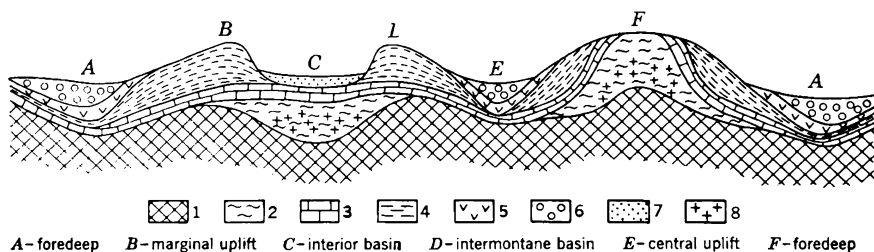


FIG. 149. Diagrammatic section showing the distribution of sequences of a geosyncline at the end of a cycle (vertical scale greatly exaggerated): (1) ancient basement; (2) lower terrigenous sequence; (3) limestone sequence; (4) upper terrigenous sequence; (5) lagoonal sequence; (6) molasse sequence; (7) interior basin sequence; (8) ophiolites.

In the present central uplift the structure is that of a great anticlinorium. On the erosional surface, one finds a succession of rock outcrops corresponding to the order of their formation: the oldest preinversion sequences occur in the axial part of the anticlinorium (lower terrigenous sequence), and the later limestone sequence is found on the flanks of the structure.

In the marginal uplifts we find sediments of the marginal deep: the upper terrigenous sequence. The present uplifts were occupied by marginal deeps immediately after the inversion. Later the deeps were incorporated into the central uplift, which was then divided into two marginal uplifts through the formation of the interior basin. Upper terrigenous sequences are also found on the flanks of the central uplift in the right half of the geosyncline. It may be expected that the fore-

¹ This geosyncline consisted of two intrageosynclines. One of them developed by forming an interior basin and dividing the central uplift into two marginal uplifts; the other developed more simply with the formation of a single central uplift.

deep sediments will be underlain by preinversion sequences distinguished by very thin sediments and very coarse facies.

In the foredeeps and intermontane basins one may observe the development of younger sediments: the lagoonal and molasse sequences. Again one can expect these to be underlain by marginal deep and even preinversion sequences, both distinguished by very thin deposits and coarse facies. An exception is the case where a foredeep or intermontane basin coincides with a marginal deep, as in the Rion depression (in the Transcaucasus). In such cases, thick-bedded upper terrigenous sequences underlie the intermontane basin sequences. Foredeeps and intermontane basins are usually underlain only by the upper members of preinversion sequences, because the site of these deeps before the inversion was occupied by intrageanticlines or the edges of platforms that subsided and collected sediments only toward the end of the first stage of the cycle.

The Greater Caucasus is an example of regular disposition of sequences. The axial part of the range has outcrops of Jurassic and Lower Cretaceous lower preinversion beds, which on the periphery of the range are bordered by outcrops of the limestone sequence dating from the Late Cretaceous and the Eocene. Farther on, on the flanks of the anticlinorium, there are outcrops of Maykopian beds, which represent the upper terrigenous sequence of the marginal deep. Finally, the adjoining foredeep and intermontane basins contain typical Neogene and Quaternary molasse sequences.

Parageosynclines are a special case. Since they show no evidence of local inversion, all sequences accumulated in the course of a cycle are found in one place. The character of these sequences depends on the history of the given parageosyncline. In the Subalpine parageosyncline, for example, preinversion sequences are overlain by foredeep sequences.

From the principles thus far established, one can forecast the character and thickness of the section if one knows the tectonic zone involved. On the other hand, the association of certain sequences with certain tectonic zones of the geosyncline is so obvious that this association can be used to determine the zone in question. This approach is especially useful in studying ancient geosynclines that have been affected by the crustal movements of subsequent cycles. From the sequences one can determine from which part of the old geosyncline the new geosyncline was formed. For example, the fact that in the Greater Caucasus the Permian sediments are represented by a red-colored molasse shows that this Alpine intrageosyncline was formed on the site of a Hercynian intermontane basin.

In delimiting tectonic regions on the basis of their sequences, however, one must bear in mind that, first, various sequences in various cases may appear in different stages of completeness and, second, the

distribution of the sequences may have been complicated by particular, superimposed cycles. The succession of preinversion beds in the Greater Caucasus will serve as an example: a terrigenous series of the Lower and Middle Jurassic, a predominantly limestone series of the Upper Jurassic, a terrigenous series of the Lower Cretaceous, and a limestone and marl series of the Upper Cretaceous and Eocene. If it is assumed that these have been produced by a first-order cycle, then all deposits from the Lower Jurassic to the Lower Cretaceous inclusive belong to the lower terrigenous sequence (1), and the rocks of the Upper Cretaceous and Eocene belong to the limestone sequence (2). The appearance of limestones in the Upper Jurassic must be attributed to a secondary cycle.

Complications also arise in the succession of sequences in the Urals. Here the Hercynian cycle begins with the Lower Devonian. On the western slopes of the Urals, this stage is represented mainly by continental clastic deposits. Limestones predominate in the Middle and Upper Devonian and are most widely developed in the Upper Viséan stage. But this thick limestone sequence of the Devonian and Lower Carboniferous is interrupted by coal-bearing strata of the Tournaisian and Lower Viséan stages, which correspond to a secondary cycle.

It follows from the preceding discussion that the various terrigenous formations may obtain their material from different sources. Preinversion sequences are composed of clastic material from platforms and intra-geanticlines. The coarseness of these sediments therefore increases toward the foredeeps and intermontane basins, if one considers this phenomenon within the framework of the interior structure of the geosyncline at the final stage of its development. The coarseness of the sediments of the marginal and intermontane basins and foredeeps, on the other hand, increases toward the central uplifts or the interior basins, since these zones were subjected to erosion immediately after the inversion. On the southern slopes of the Greater Caucasus, for example, the coarseness of the Mesozoic deposits increases southward in the direction of the Kura depression, where uplift preceded the inversion. The Tertiary deposits of the depression, on the other hand, become coarser toward the main mountain range.

We must now consider the special conditions under which the marginal deep (or upper terrigenous) sequence accumulates. This sequence shows great uniformity of facies both in section and in plan. The coal measures of the Donets basin were cited above as an example of this sequence; other coal-bearing series of the same age are similar. Excellent examples of facies uniformity are seen in the Maykopian series of the Greater Caucasus, in the graptolite shales in Caledonian geosynclines, and in the flysch deposits of the Carpathians, the Alps, and many other geosynclinal regions. For this uniformity to appear in the

vertical section (through long deposition of the same type of rock), the uplifts that supplied the clastic material and the depressions where it accumulated had to be of uniform amplitude. Only under such conditions could the deposition exactly match the subsidence and the surface of the sediments be maintained on the same level.

Since subsidence predominates before the inversion and uplift after the inversion, there must be an intermediate stage in which these movements are in balance. Such a stage would have to occur at the time of inversion. Judging from the fact that facies maintain their uniformity for some time after the inversion (in the Donets basin during the Middle and Late Carboniferous, in the Caucasus during the Middle and Late Oligocene, and elsewhere even later), the stage of balanced crustal movements must last for a considerable period before uplift finally becomes predominant.

The uniformity of the facies in the horizontal section is due to other factors. At the time of inversion the basic forms of surface relief are reversed: a low-lying region of deposition becomes a zone of uplift and erosion, and areas of erosion become zones of deposition. These fundamental changes in local topography apparently do not occur suddenly but in gradual stages. Between one extreme state and the other, the locality passes through an unstable intermediate stage while the old distribution of erosional and depositional zones is being destroyed, but the new distribution is not yet stabilized. During this transitional period, the great oscillatory movements of the earth's crust that determine the position of the shore lines lose their stability, and the shore lines become extremely mobile. Zones of erosion and deposition arise here and there, constantly changing places, while the shore lines are displaced over great distances; this creates conditions for a long period in which the sediments are washed and the facies dislocated by wave action. As a result, a sedimentary sequence dating from the time of the inversion displays an extraordinary uniformity of facies.

Why are these periods of balanced uplift and subsidence and of uniform facies favorable for the accumulation of coal and bitumens in the sediments? This question still cannot be answered fully. It may be supposed that while the bottom surface of a basin remains at the same hypsometric level, as accumulation exactly compensates for subsidence, minor crustal movements would constantly dissect the bottom of the basin into a large number of closed or half-closed depressions. Near the shore, such depressions become lagoons in which the coal-forming process takes place. At some distance from the shore, these depressions become submarine troughs in which the water becomes stagnant and contaminated with hydrogen sulfide, thus favoring the accumulation of organic matter and its conversion into bitumens. A major factor in the accumulation of this material is the high rate of subsidence in foredeeps

and the rapid rate at which the organic matter is buried by subsequent deposition. Climate is probably also important. The time of the inversion coincides with the formation of extensive seas and of a low relief, which, in turn, produce an equable humid climate favorable to the growth of vegetation, especially in an amphibious and swampy environment.

Thus one can also understand the appearance of uniform coal-bearing strata at the beginning of the cycle. At that time, the crustal movements in geosynclines again pass through a stage in which uplift and subsidence are in balance, but this time in the opposite direction, passing from a predominance of uplift to a predominance of subsidence. The cycle begins with the great uplift that ends the preceding cycle. Uplift thus predominates at the very start of the cycle. But subsidence then increases in intensity, becoming predominant by the end of the first half of the cycle. In the interval the cycle must pass through a neutral stage.

SEDIMENTARY SEQUENCES OF PLATFORMS

There is a widespread opinion that the facies of geosynclinal sediments differ significantly from platform facies because of the greater depth of geosynclinal marine basins and their more open nature. Careful study of this problem shows that the difference has been greatly exaggerated. There is no evidence that marine basins in general are deeper in geosynclines than on platforms. The great subsidence in geosynclines is compensated by the rapid accumulation of sediments, so that geosynclinal sections as a rule contain both sediments of shallow-water origin and continental deposits.

The concept of facies is often used, consciously or unconsciously, to include the thickness of the deposits. In that case, of course, there would be a considerable difference between the accumulation of sediments in a geosyncline and that on a platform. But the facies of deposits must not be confused with thicknesses. The thickness of the deposits itself gives no indication whatever of the depth of the basin. It shows only the total amount of subsidence of the earth's crust that has been compensated by the accumulation of sediments. The same sands, clays, and limestones of various types that are found in geosynclines also occur on a platform. Rocks of geosynclinal and of platform origin differ very slightly or, perhaps in some areas, not at all.

One cannot, however, completely deny that there are differences. Since the subsidence in a geosyncline is generally of longer duration than on a platform, marine sediments, on the whole, may be expected to predominate in geosynclines. Moreover, certain subsequences are found only in geosynclines: the flysch subsequence is such a one. On

platforms, because of less intensity of uplift and erosion, there are no coarse clastic deposits such as coarse conglomerates or boulder-sized accumulations, whereas in geosynclines, at the foot of uplifts, very coarse material accumulates together with molasse deposits. Among the sandstones, quartz material predominates on platforms and graywackes are widely developed in geosynclines. Platform deposits rarely show any traces of submarine slides.

With these and certain other exceptions, platform sequences are similar to geosynclinal sequences and accumulate in the same succession during the course of a cycle. The Russian platform may serve as an example. The history of this platform falls into three cycles. The first, or Caledonian, cycle begins in the Baltic region with the accumulation of the Lower Cambrian terrigenous deposits (lower terrigenous sequence, 1). The Lower Silurian is represented largely by limestones (limestone sequence, 2). Dolomites predominate in the Upper Silurian (lagoonal sequence, 4). Finally, the Devonian is represented by red-colored continental deposits (molasse sequence, 5). In view of the fact that graptolite shales are widely developed in the Silurian of Sweden, in an area forming part of the same platform, the upper terrigenous sequence (3) can be added to this list.

The Hercynian cycle, in the central regions of the European part of the U.S.S.R., begins with the appearance of terrigenous deposits of the Middle Devonian (sequence 1), which are soon replaced by limestones (sequence 2). The Lower Carboniferous shows mainly coal- and oil-bearing facies (sequence 3), which are once again overlain by limestones. This particular succession must be ascribed to the effect of a secondary cycle. In the Permian system there are many hydrochemical deposits (lagoonal sequence, 4), and the cycle concludes with an accumulation of the red-colored Tatarian deposits (molasse sequence, 5).

The Alpine cycle also begins with the accumulation of terrigenous deposits, of the Jurassic and the Lower Cretaceous (sequence 1), followed by the limestones and marls of the Upper Cretaceous (sequence 2). After this, terrigenous deposits again predominate. The Quaternary period produced salt deposits in lakes and lagoons (sequence 4) and an accumulation, at the foot of the Caucasus, of various clastic rocks, which not only fill the foredeep of the Caucasus geosyncline but extend far into the platform (sequence 5).

This example shows that each platform cycle contains essentially the same succession of sediments as the geosynclinal cycle, except that the flysch subsequence is missing from the platform succession. As far as the mineral- and fuel-bearing subsequence is concerned, it should be noted that this not only occurs on platforms, but it accumulates there at the same stage at which it appears in neighboring geosynclines. In the central parts of European Russia, for example, these sequences are

of the same age as in the adjoining Hercynian geosyncline of the Urals. Similarly, the oil-bearing Pennsylvanian deposits in the United States are found far to the west of the Appalachians, within the platform.

It has been seen that, except for the inversion stage, the beginning of the cycle is the most favorable time for the accumulation of organic matter in geosynclinal sediments. Platform sediments analogous in time and conditions of accumulation are also found. For example, oil-bearing and bituminous deposits are encountered on the Russian platform in the Middle Devonian sediments associated with the very beginning of the Hercynian cycle. The beginning of the Alpine cycle is represented by the oil-bearing Jurassic deposits of the Ural-Emba region. Evidently the appearance of rocks rich in organic matter has been determined, on platforms as in geosynclines, by the balance between uplift and subsidence, which occurs twice during the cycle: the first time during the transition from predominant uplift to predominant subsidence, and the second time during the reverse transition.

Since no local inversions occur on platforms, all sequences, without exception, accumulate in subgeosynclines and, to less extent, in subgeanticlines, overlying each other.

It should be noted that the chief sources of the clastic material deposited on platforms must be sought outside the platforms. Such sources may be found in the mountain ranges that have been uplifted in the adjoining geosynclines. The most abundant source of material is always to be found in a very young uplift rising in a geosyncline that has completed its development during the preceding cycle. In the Hercynian cycle, for example, although the Urals did not yet exist, the Scandinavian mountains, which had been uplifted at the end of the Caledonian cycle, were the chief source of clastic material for the Russian platform. The Urals became such a source toward the end of the Hercynian cycle and retained that role through almost the entire Alpine cycle. It was only toward the very end of the Alpine cycle that the Caucasus Mountains became the chief source of clastic sediments. Interior zones of erosion situated in subgeosynclines usually contributed very little material, since the degree of uplift was very small. As in geosynclines, the succession of sequences on platforms may be complicated by secondary, superimposed cycles.

GENERAL CONCLUSIONS

The preceding discussion shows clearly that sedimentary facies and sequences are closely related to the oscillatory movements of the earth's crust. Facies and sedimentary sequences appear in a succession dictated by the orderly process of these movements, and their distribution in both time and space is also determined by the laws of that process.

Therefore it should now be possible to explain the observed irregularity in the distribution of known coal and petroleum deposits in the stratigraphic column. Figure 150 shows a curve of the distribution of the world's coal reserves in the deposits of various systems. The maxima on the curve in the Carboniferous and Tertiary periods correspond to inversions and the formation of foredeeps in the Hercynian and Alpine cycles. A smaller maximum in the Jurassic corresponds to the beginning of the Alpine cycle. Figure 151 shows a similar curve for petroleum. It resembles the curve in Fig. 150, since the accumulation of bitumens takes place in the same stages of the cycle as the accumulation of coals.

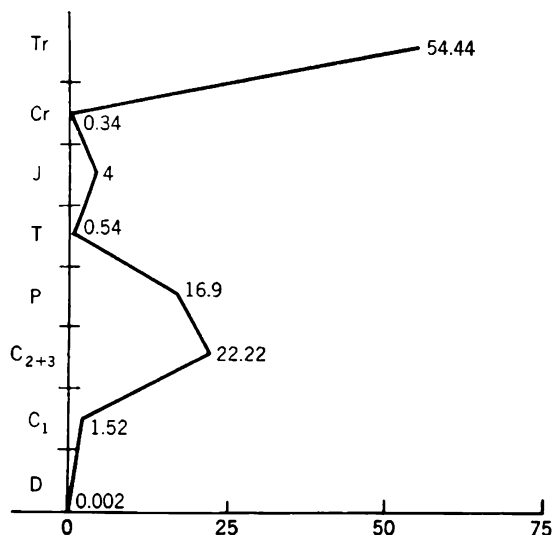


FIG. 150. Distribution of the world's coal reserves by geologic systems (after P. I. Stepanov).

N. M. Strakhov (1949) has studied the periodicity in the distribution of a number of other sedimentary minerals (bauxite, iron, manganese, phosphate, gypsum, and salt) and has established a close association between epochs of maximum accumulation and certain stages of the tectonic cycle.

The regularity of the accumulation of sedimentary sequences is related to the periodic nature of the oscillatory movements. But this periodicity is superimposed on a progressive development. We must therefore conclude that, in addition to the periodic repetition of analogous sequences, there must be differences between cycles and that these differences increase with time. In other words, there must be some indications of irreversibility in the development of sedimentary deposition, just as there is irreversibility in the development of the oscillatory movements. Unfortunately, this aspect of the earth's geologic history

has been insufficiently studied. The first attempt in this direction was made by N. M. Strakhov (1949), who found that certain rocks are quite characteristic of Precambrian times but almost entirely lacking later and that, on the other hand, other rocks predominate in post-Cambrian times. Rocks of the first category include jaspilites, some manganese ores of marine origin, limestones, and siliceous rocks of chemical origin. Rocks of the second category include coals, salts, bog iron and oolitic iron ore, bauxite, limestones, and siliceous rocks of organic origin.

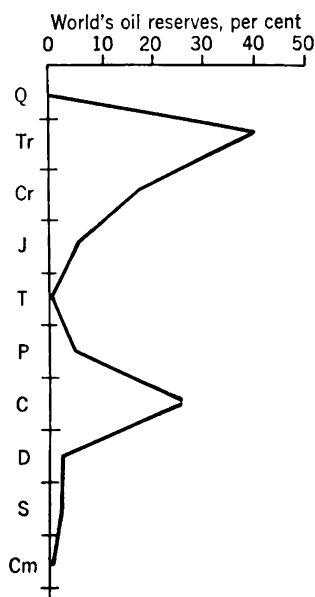


FIG. 151. Distribution of the world's petroleum deposits by geologic systems (after I. M. Gubkin).

The reasons for the irreversibility of the process of formation of sediments must be sought mainly in the irreversible evolution of the organic world and in the irreversibility of tectonic development, which up to now has led to an expansion of the platforms at the expense of the geosynclines. The first factor accounts for changes in the role of organisms in the formation of sediments. In view of the great importance of living organisms in the current processes of sedimentation in all its stages, one can easily understand that these processes must have operated quite differently in ancient times, when life on the earth was either absent altogether or just beginning. The second factor accounts for the decrease in the relative abundance of coarse clastic rocks, the leveling out of the differences in facies, and the increasing development of clay facies.

This discussion has been limited to sedimentary sequences. Actually the term *sequence* is much broader and refers to regular complexes of both sedimentary and igneous rocks. Sequences also differ sometimes in the degree of metamorphism of their rocks. In a broader sense, therefore, a sequence includes all the rocks characteristic of a given stage of a cycle in a given geotectonic zone, both sedimentary and igneous, as well as the metamorphic changes that these rocks subsequently undergo in a regular manner because of their location in zones of certain movements, intrusive activity, and so forth. This broader meaning of sequences will be discussed later (in Chap. 35).

PART FIVE

TECTONIC MOVEMENTS THAT PRODUCE FOLDING

CHAPTER 18

BASIC PRINCIPLES OF THE DEFORMATION OF SOLID BODIES

In Parts Five and Six are discussed the mechanical processes that occur in rocks as the result of tectonic forces. Such processes are manifested, in the first place, in deformations—in changes in the shapes or volumes of geologic bodies—and, in the second place, in the rupturing of the rocks through the formation of tectonic fractures or faults.

Tectonic forces may produce both elastic and plastic deformation in rocks. Only irreversible plastic deformations, however, are permanently marked in the structure of the earth's crust. Folding—the most common expression of deformation in the rocks—is the most important manifestation of their plastic deformation.

Before going into the particular problems of tectonics, one must have some brief and elementary acquaintance with the modern theory of deformation of solid bodies, which include rocks. This theory, however, which is necessary for a proper understanding of the conditions under which deformations appear in the earth's crust, is far from sufficient for the solution of all the tectonic and physical problems that arise in connection with the study of folded structures. The reason is that the theory of plastic deformation, as worked out up to now, is far from complete. In physics and engineering, the theory of plastic deformation has been worked out almost exclusively in connection with particular practical problems, mainly as applied to metals; but metals, it is well known, differ considerably in their structure and properties from rocks.

Another important aspect is the fact that in physics and engineering only deformations of relatively small magnitude are studied. Under geologic conditions, however, rocks are frequently subjected to deformation on an enormous scale, associated with very great relative displacements of individual masses of rock.

The lack of uniformity of the deformed bodies is of great importance in tectonic deformation. The lack of uniformity of sedimentary rocks, as reflected in their bedding, for example, is the fundamental reason why the plastic deformation of such bodies takes the form of folding.

Physics and engineering, on the other hand, rarely study deformations of nonhomogeneous bodies, and then only as applied to special cases that have little in common with tectonic conditions.

Mention must also be made of the difficulty of determining the external forces that have produced deformations in rocks under the conditions in the earth's crust: the geologist observes only the results of the deformation and is compelled to infer the nature of the deforming forces from them. Another essential aspect is the variety of factors affecting the process of deformation under geologic conditions (duration of the deformation, temperature, changes in the properties of the rocks produced by moisture, changes in the internal structure of the rocks in the course of the deformation, etc.).

All these remarks will indicate that, in the present state of the problem, it is impossible to base any description or classification of folding on a single sufficiently rigorous and detailed theory of plastic deformation. Such a theory has not yet been elaborated even for the very simplest cases. Particular parts of the modern theory of deformation can, however, be used in solving tectonic problems. Such aspects will be presented in the present chapter, the material of which is derived from various sources in scientific literature (Gubkin, 1946, 1947; Ivanov, 1948; Peshl', 1948; Uzhik, 1950; Filonenko-Borodich et al., 1949; Fridman, 1946; Shreyner, 1950; and others).

The further development of these conceptions of the mechanism of tectonic deformations requires much special experimental and theoretical research; this is the subject of a new branch of science called *tectonophysics*.

STRESSES AND STRAINS

A solid body subjected to a load—that is, to the action of mechanical forces balanced in such a manner that they do not produce any displacement or rotational movement of the body—undergoes deformation involving a change in its shape, its volume, or both. Depending on the directions in which the forces are applied, the physical body will be subjected to various stresses, the simplest varieties of which are *compression*, *tension*, *shear*, *bending*, and *torsion*, and will undergo corresponding deformation (Fig. 152).

In the case of an elementary or unit body of infinitely small volume, however, the theory of elasticity states that any combination of complex deforming forces applied to such a body in any directions (if the body is isotropic) will result only in the action of compressive or tensile forces in three mutually perpendicular directions and will produce compressive or tensile strains in these directions. These directions are called the *principal strain axes*. This conclusion in the theory of elas-

ticity enables us to consider only the action of the forces of compression and tension; it must be kept in mind, however, that deformation in compression or tension is always indissolubly connected with shear, inasmuch as the process of compression and tension always involves changes in the angles between imaginary lines drawn through the body undergoing deformation.

The deformation is uniform if all parts of the body are deformed in the same manner and to the same degree and if the principal strain axes are everywhere in the same direction, and it is nonuniform if the nature and extent of the deformation change from one part of the body to another and if the principal strain axes change their directions within the body. Examples of nonuniform deformation are bending and torsion (Fig. 152*d,e*). Bending, to the first approximation, may be regarded as tension and compression differing in different parts of the

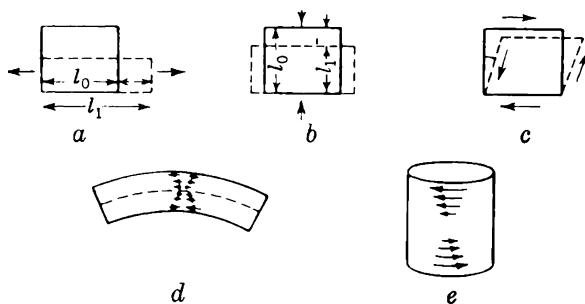


FIG. 152. Various types of deformation: (a) tension, (b) compression, (c) shear, (d) bending, (e) torsion.

body. Torsion, to the first approximation, is shear differing in magnitude at different distances from the axis of the twisted cylinder (Fig. 152*e*). In other cases, one may observe combinations of nonuniform compression or tension and shear.

Deformation in tension and compression is characterized by the magnitude of the relative elongation ϵ , which is taken to be positive in tension and negative in compression (Fig. 152*a,b*). Hence,

$$\epsilon = \frac{l_1 - l_0}{l_0} = \frac{\lambda}{l_0}$$

where l_0 = the original length of the body

l_1 = the length of the body after its deformation

λ = absolute elongation

When the deformation is uniform, the magnitude of ϵ is the same at all points.

Shear is defined as the magnitude of the change in any original right

angle in a prismatic body; in small deformations this magnitude may be replaced by the tangent of the angle of shear (Fig. 152c).

When the stresses increase up to a certain limit, the solid body, as a rule, undergoes elastic deformation. Such deformation is reversible: when the deforming force is removed, the deformation also disappears and the body returns to its original shape. With a further increase in the deforming force, the elastic deformation, in the general case, becomes plastic, or permanent. Such deformation does not disappear when the deforming force is removed. With a still greater increase in the force of deformation, there will be a point at which the body undergoes rupture by the formation of one or more fractures in it.

The solid body exerts a resistance against the deforming forces, which they must overcome in the process of deformation. This means that, in

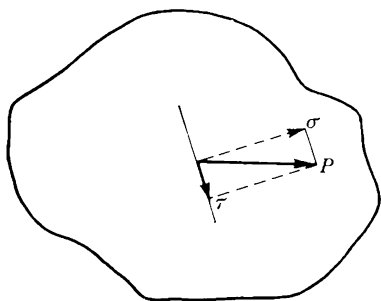


FIG. 153. Total (P), normal (σ), and tangential (τ) stresses.

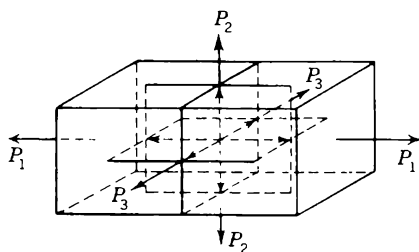


FIG. 154. Diagram showing extension of a parallelepiped in three mutually perpendicular directions.

the process of deformation, forces arise in the body that oppose the external forces. These internal forces, when expressed in terms of a unit area of some imaginary section drawn through the body, are called *stresses* acting on the given unit area. The total stress acting on the given unit area may be resolved into two components: a stress acting perpendicular to the unit area (the normal stress) and a stress acting within the plane of the unit area (the tangential stress) (Fig. 153).

In the case of an isotropic body, the normal and tangential stresses acting on any unit area at a given point determine the stresses in this area. If the isotropic body undergoes uniform deformation, the stresses at all points of the body will be the same and the stress state will thus be uniform. If, however, the stresses change from point to point, it will be nonuniform. The following examples, for the sake of simplicity, will always involve stresses arising from the uniform deformation of an isotropic body. •

In a given deformation of a body, the relationship between the normal and tangential stresses will differ, depending on the relative orienta-

tion of the elementary or unit area under consideration. When the body undergoes tension, for example, if the unit area is perpendicular to the axis of tension, the normal stresses will have the greatest magnitude and the tangential stresses in this case will be zero. In the planes or unit areas parallel to this axis, there will be neither normal nor tangential stresses; in the planes with intermediate attitudes, normal and tangential stresses will both be present but in different relationships to each other.

The theory of elasticity states that in any deformation of a body there will be three mutually perpendicular planes, across which there are only normal stresses and in which the tangential stresses are absent. These three normal stresses together fully characterize all the stresses acting on a point: these are the *principal stresses*, and their directions are called the *principal stress axes*. In an isotropic body they are at the same time the principal strain axes.

In nonuniform deformation, the directions of the principal axes change from point to point. In an isotropic body undergoing uniform deformation, the principal axes are mutually parallel at all points, and the stresses at any given point are therefore fully characterized by a single system of three principal stresses.

The location of the principal axes is of great importance: if a system of coordinates is selected such that the coordinate axes coincide with the principal axes and all the forces acting along these three axes are of equal magnitude, the stress state is much more simply characterized.

Figure 154 shows an isotropic body subjected to uniform extension in three mutually perpendicular directions. The directions of the forces shown, which are normal to the faces of the parallelepiped, coincide with the principal axes: the unit areas in the planes perpendicular to them are acted upon only by normal stresses, which fully characterize the stress state existing in the body. If we were to select some other planes, the tangential stresses would also have to be included in the characterization of the stress state.

In the general case, the principal normal stresses are of different magnitudes, so that one may distinguish a greatest, an intermediate, and a least principal stress. In addition, tensile stresses are considered to be positive and compressive stresses negative. One or two of the principal stresses may be equal to zero. If only one of the principal stresses is not equal to zero, the stress state is *uniaxial*, or *linear*, as, for example, in the longitudinal extension of a thin wire. If only one of the principal stresses is equal to zero, the stress state is *biaxial*. An example is the deformation of a thin plate by forces lying in its plane. If a right prism is acted upon by stresses perpendicular to one of the edges and constant along all the surfaces of the prism, the resulting stress state may also be considered biaxial. In that case, instead of the prism, let us consider

a thin plate torn apart perpendicular to the edge, as long as the stress state at any point, regardless of the part of the plate from which it is taken, is the same. This is the typical case, in which the three-dimensional problem may be reduced to a plane problem. Later on, in analyzing the mechanism of formation of tectonic structures, we shall, for simplicity, be dealing mainly with the plane problem. If, finally, none of the three principal stresses is equal to zero, the stress state is *three-dimensional*, or *triaxial*.

Let us consider the relationship between the normal and tangential stresses as a function of the orientation of the unit area in various states of stress in the body. All these cases will involve uniform stresses and small deformations of bodies that are right parallelepipeds. We may begin with uniform stress in which there is only one principal stress, expressed as tension or compression along a single axis (Fig. 155). This is approximately the case in the extension of a fine wire.

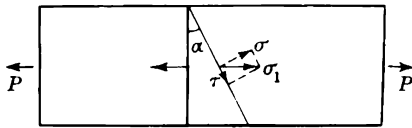


FIG. 155. Diagram of normal and tangential stresses in tension.

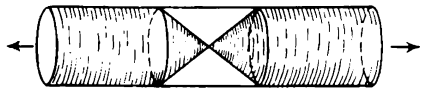


FIG. 156. Cone formed by the maximal tangential stresses in uniaxial tension in a cylinder.

Let σ be the normal stress on any given unit area, σ_1 the normal stress in a cross section through the body perpendicular to the tensile or compressive stress, τ the tangential stress in any plane unit area, and α the angle between the given unit area and the perpendicular cross section. Then the magnitude of the normal stress in any unit area, in the case of tension, will be given by the formula

$$\sigma = \frac{\sigma_1}{2} (1 + \cos 2\alpha),$$

and the tangential stress will be equal to

$$\tau = -\frac{\sigma_1}{2} \sin 2\alpha$$

From these formulas, it follows that the maximal normal stress $\sigma_{\max}$ occurs when $\alpha = 0$ or that the normal stress in uniform extension will be greatest in the cross section, so that $\sigma_{\max} = \sigma_1$.

In the same unit area the tangential stresses are equal to zero. Thus the maximal normal stress $\sigma_{\max}$ is the principal stress acting in the isotropic body along the principal strain axis.

From the same formulas it follows that the tangential stress will be greatest when $\alpha = 45^\circ$; that is, in the plane unit area inclined at an

angle of 45° to the principal axis. The theory of elasticity establishes the law of the relationship between the tangential stresses according to which the tangential stresses in two planes, forming an angle of 90° between them, are equal in magnitude. Hence the greatest tangential stresses will be observed in the planes inclined at an angle of 45° to the axis of tension. In the case of a uniaxial stress state, these planes in combination will form a double cone (Fig. 156).

The magnitude of the greatest tangential stress in uniaxial tension is related to the magnitude of the greatest normal stress by the following equation:

$$\tau_{\max} = \frac{\sigma_{\max}}{2}$$

In uniaxial compression, the positions of the plane unit areas with the greatest normal and tangential stresses will be the same, but in the formulas the normal stresses must be designated by a minus sign.

Biaxial stresses may have a single sign, if both the principal stresses are either positive or negative, or different signs, if one of the principal stresses is compressive and the other is tensile. Let us assume that we have a biaxial positive (tensile) stress state with a single sign (Fig. 157). Let the principal stresses be designated as σ_1 and σ_2 , and let $\sigma_1 > \sigma_2$. The greatest normal stress will be observed in the plane unit areas perpendicular to the direction of the maximal tension (the axis of maximal tension), and it will be equal to σ_1 . The greatest tangential stress will be observed in the planes that bisect the angles between the areas of the principal normal stresses; that is, in the planes that form an angle of 45° with the principal axes. In this case the combination of all such areas will form an intersecting plane.

The magnitude of the greatest tangential stress in biaxial tension is determined by the formula

$$\tau_{\max} = \frac{\sigma_1 - \sigma_2}{2}$$

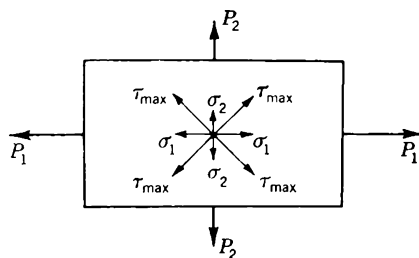


FIG. 157. Distribution of the principal and maximal tangential stresses in biaxial tension.

Compressive stresses are considered negative; the position of the plane unit areas with the extreme values of the normal and tangential stresses in this case remains the same.

From the last formula, it may be seen that if both principal stresses are equal, there will be no shear stresses. There are no shear stresses, for example, in a plate subjected to equal compression along two

mutually perpendicular axes lying in the plane of the plate. For tangential stresses to arise, the principal normal stresses must differ either in magnitude or in sign.

Triaxial stress states are depicted in Fig. 158. The three principal stresses are designated as σ_1 , σ_2 , and σ_3 . Let $\sigma_1 > \sigma_2 > \sigma_3$, so that σ_1 is the greatest principal stress, σ_2 is the intermediate principal stress, and σ_3 is the least principal stress. In this case, the maximal normal stresses will evidently be observed in planes perpendicular to the greatest principal stress σ_1 . The maximal tangential stress will be found in the two

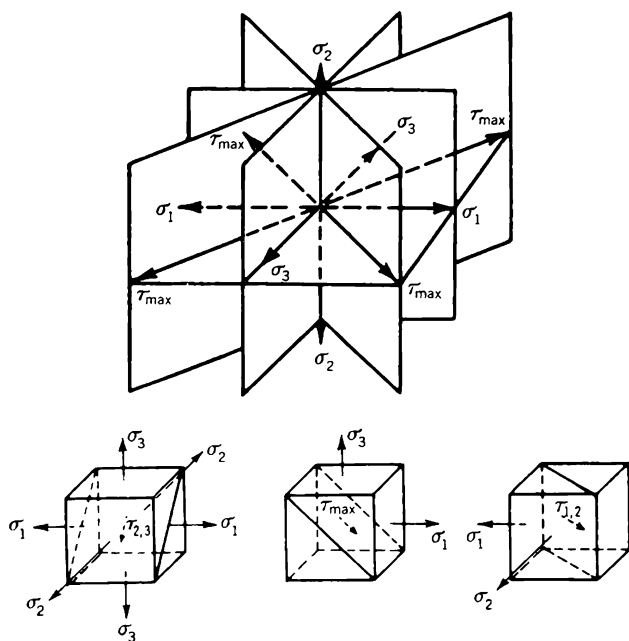


FIG. 158. *Upper*, principal and maximal tangential stresses in triaxial tension; *Lower*, tangential stresses produced by various normal stress couples.

mutually perpendicular planes that bisect the right angles between the planes with the maximal and minimal magnitudes of the principal normal stresses. In this case, the magnitude of the maximal tangential stress will be

$$\tau_{max} = \frac{\sigma_1 - \sigma_3}{2}$$

In the planes that equally divide the other angles between the planes with the principal stresses, the tangential stresses will be the following (Fig. 158):

$$\tau_{1,2} = \frac{\sigma_1 - \sigma_2}{2}; \quad \tau_{2,3} = \frac{\sigma_2 - \sigma_3}{2}$$

From the foregoing, it is clear that tangential stresses arise only when the principal normal stresses have different signs or different magnitudes. When any two of the principal stresses are equal, the tangential stresses that derive from them will be zero; in the case of ambient compression or tension acting in all directions, when all the principal normal stresses are equal, there are no tangential stresses.

This section may conclude with a brief discussion of the means of determining the positions of the principal stress axes in studying structures of tectonic origin. In many cases, such a determination will not encounter great difficulties, if the deformed elements of the rock may, with sufficient degree of approximation, be considered isotropic and if their principal stress and strain axes may be identified. For example, the flattening of the pebbles and grains in a rock, as well as the orientation of its minerals, determined macro- and microscopically, may indicate the positions of the principal axes of tension and compression. If the rock cannot be considered isotropic, the determination will be more difficult. On the other hand, when the individual anisotropic elements of the rock are randomly disposed and are sufficiently numerous, it has been established statistically that the average positions of the strain axes are close to the positions of the stress axes. But if the anisotropic elements (for example, crystals) are initially oriented, the principal strain axes will not coincide with the principal stress axes and will be extremely difficult to determine.

When observations show that the rock has undergone neither compression nor tension but has been subjected to some other action ("load"), difficulties of another kind are encountered. An extremely widespread type of tectonic "load" is shear stresses, which in the simplest case form a couple. Figure 159 shows a rectangular plate subjected to such forces. This plate may represent any given block of rocks located between two parts of the earth's crust moving parallel to each other but in opposite directions (such as part of a bed of rocks between two other beds moving parallel to the bedding planes in opposite directions). In such cases the usual diagram shows a single couple. But inasmuch as the body subjected to these forces must be in equilibrium—that is, it cannot either move or be rotated in space—there must also be a second couple, which prevents the rotation of the body that would be produced by the first couple alone.

Adding together the couples, as indicated in Fig. 159, in the case under consideration we obtain two principal axes, one of which is the axis of tension and the other the axis of compression. These axes are at an angle of 45° to the direction of the forces.

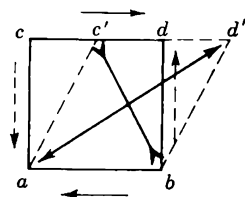


FIG. 159. Shear in a thin plate: ($a-d'$) axis of greatest elongation; ($c'-b$) axis of greatest shortening.

It must be stressed that up to this point we have been dealing with uniform deformation of isotropic bodies. Under these conditions, one may assume that the stresses are also uniform and do not change noticeably either in space or in time. Under geologic conditions, however, one is always dealing with anisotropic bodies and nonuniform deformations. This results in nonuniform stresses and in complex distributions of the stresses in space. The plastic deformations, as already indicated, are usually large under geologic conditions. When added to the changes in the forces applied to the rocks and to the changes in the properties of the deformed material during the process of deformation, they may lead to changes in the stress not only in space but also in time. In addition, the positions of the principal strain axes are also subject to constant change.

These circumstances introduce great complications into the mechanical interpretation of the geologic structures observed in the earth's crust. For a complete mechanical analysis of such structures, it would be necessary to determine the positions of the principal stress axes for each infinitesimal part of the rock individually and to calculate the changes in the positions of these axes during the course of the deformation.

It was said above that as the forces of deformation increase, the deformation first increases but remains elastic. If one imagines the solid body to be composed of infinitesimal hinges joined by springs, in such a model the deformation would take the form of a displacement of the hinges relative to each other accompanied by changes in the tension of the springs. In the directions of the principal normal stresses, the hinges are either closer to each other or farther removed (in certain directions they may be absent), but they will also be displaced laterally relative to each other; that is, in these directions, the lines joining the hinges will be rotated. In other words, along the principal axes the body is merely elongated or shortened, whereas in the other directions it is also sheared. In addition, the particles undergoing the greatest shear in isotropic bodies will be along the directions of the greatest tangential stresses.

The relationship between the stresses and the magnitude of the elastic deformation, as long as the latter is still small, is determined with sufficient accuracy by Hooke's Law, or the law of proportionality. In the case of uniaxial tension, this is

$$\sigma = E\epsilon$$

where σ = stress

ϵ = magnitude of the deformation

The coefficient of proportionality E is called the *modulus of elasticity*, or *Young's modulus*.

Similarly, the shear modulus is

$$\tau = g\gamma$$

where τ = tangential stress

γ = magnitude of shear

g = shear modulus

In the case of ambient (hydrostatic) compression, one may speak of the *volumetric compression modulus*.

Elastic tension and compression are accompanied by changes in volume. Experiments have shown that if a body is subjected to uniaxial tension, it will undergo compression in the directions perpendicular to the tension, but this compression will be of somewhat smaller magnitude than that required to maintain the original volume of the body. Similarly, in uniaxial compression there is lateral expansion that does not fully compensate the compression. Consequently, in elastic tension the volume of the body is always increased, and in elastic compression it is decreased. The magnitude of the change in the lateral dimensions of the body differs for different materials. If the lateral diameter before deformation was equal to d_0 and after deformation is equal to d_1 , the relative decrease in this lateral dimension will be

$$\epsilon_g = \frac{d_1 - d_0}{d_0} = \frac{\Delta d}{d_0}$$

The relationship between the relative change in the lateral dimensions and the relative change in the longitudinal dimensions is called *Poisson's ratio*. The magnitude of Poisson's ratio is always less than 0.5. For the majority of materials it is between 0.333 and 0.250.

An ideal elastic body must be deformed proportionally to the force applied to it almost instantaneously (at the velocity of the elastic waves). The deformation must also disappear with the same speed, as the force is removed. In actual bodies, one always observes an elastic aftereffect, manifested as a certain lag in the deformation, both when the load is applied and when it is removed. This is due to the fact that the deformation must overcome a certain amount of internal friction.

This discussion of elastic deformations may end with some figures (Table 13) characterizing the elastic properties of the most common rocks and minerals (Shreyner, 1950).

In the general case, as the load increases, the solid body reaches its elastic limit or limit of flow, after which plastic deformation begins. In contrast to elastic deformation, this is retained after the load has been removed. In plastic deformation (if it is sufficiently great), one may assume that the volume is constant: although the shape is changed, the volume remains practically unchanged. A change in volume can be only elastic and can occur in plastic deformation only to the extent that it is accompanied by elastic deformation. This rule does not take into

Table 13. Elastic Properties of Common Rocks and Minerals

<i>Rock or mineral</i>	<i>Young's modulus, kg/mm²</i>
Quartz.....	7,850-10,000
Feldspar.....	Up to 8,000
Calcite.....	5,800-9,000
Gypsum.....	1,200-1,500
Halite..	Up to 4,000
Granite.....	Up to 6,000
Basalt.....	Up to 9,700
Limestone.....	Up to 8,500
Quartzite.....	Up to 10,000
Sandstone.....	5,000 or more
Sand.....	30
Shale.....	1,500-2,500
Clay.....	30

account the special conditions that may exist in rocks: in compression, for example, the rocks may change their volume through a loss of water or decrease in the interstices between the grains, and such changes will be irreversible.

In the case of an ideal plastic body, the stress-strain curve after the elastic limit has been reached should become a horizontal straight line,

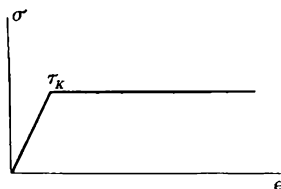


FIG. 160. Curve of plastic deformation of a body without strain hardening: ϵ , deformation; σ , stress; τ_k , elastic limit.

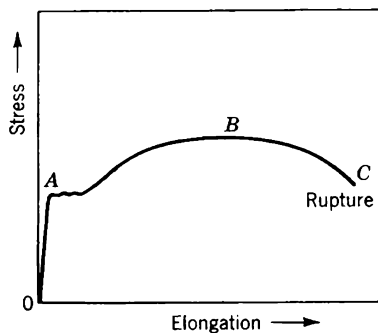


FIG. 161. General curve of deformation: O-A, elastic deformation; A-B, plastic deformation, with strain hardening; B-C, deformation in weakened body before rupture.

since in such a body a critical load may produce a deformation of any magnitude (Fig. 160). Actually, in real plastic bodies the curve beyond the elastic limit usually slopes downward after a certain time; that is, an increase in the load is required to increase the deformation. This increase in the stress, however, is considerably less than in elastic deformation (Fig. 161).

The mechanism of plastic deformation is embodied in the fact that the plastic deformation produced by a given load is accomplished

gradually and is fixed by the redistribution of the particles and by their assumption of new positions of equilibrium. In this process, the stresses that appear in the elastic deformation gradually drop to the magnitude corresponding to the elastic limit. Such "resorption" of the stresses is called *relaxation*.

The rate at which elastic deformation becomes residual deformation depends on the material and on the external conditions. In the case of liquids, the period of relaxation is very small, and the elastic deformation immediately becomes plastic and is retained to scarcely any extent. In ordinary solid bodies the time of relaxation is rather long, so that the transition from elastic to plastic deformation is far from instantaneous. That part of the deformation which changes from elastic to plastic is retained in the form of an irreversible change in the shape of the body. But even after complete relaxation, a certain elastic deformation is retained the magnitude of which corresponds to the stress equal to the elastic limit and which disappears after the load has been removed. Thus the deformation beyond the elastic limit also includes elastic deformation, which consists of two parts: one part becomes a residual deformation at a greater or less rate, and the other part is retained as elastic deformation.

If the relaxation is relatively slow and the plastic deformation consequently encounters great resistance, the body is said to have *great viscosity*. When the body has a low viscosity, the relaxation and "fixing" of the deformation take place rapidly.

Relaxation is accompanied by the phenomenon known as *creep*. This occurs when the body is subjected to stresses lower than the elastic limit and these stresses are maintained for a very long time; under these conditions (if the rate of relaxation is not too great), the redistribution of the particles owing to thermal oscillation will lead to a gradual transformation of the elastic deformation into plastic, and this plastic deformation will increase with time. The stresses will be "resorbed," and an increasingly smaller force will be required to maintain the same magnitude of deformation. A typical curve of the increase in deformation (creep) with time under a constant load is shown in Fig. 162. This curve may be divided into four parts. The first part (*A-B*) corresponds to an instantaneous or very rapid elastic deformation beginning as soon as the load is applied. The second part (*B-C*) has no independent significance and represents the transition from elastic deformation to plastic; here a slower rate of deformation is observed. The third part (*C-D*) corresponds to a stable and uniform deformation, at a constant rate depending on the viscosity of the material. The fourth part (*D-E*), finally, represents a loss of stability, the occurrence of necking in tension, and acceleration of the deformation ending at point *E* with rupture of the body.

Theoretically, any body may be deformed plastically by any small load, if sufficient time is allowed for the deformation to develop. But in a number of cases the amount of time required for any noticeable deformation to appear may be too great even in the scale of geologic time. It has also been suggested that real solid bodies, including rocks, have a *threshold of flow*, or lower plastic limit: stresses below a certain minimal magnitude cannot produce plastic deformation, no matter how long the period of time. Under geologic conditions, relaxation and creep play exceedingly important roles, inasmuch as they make it possible for enormous plastic deformations to develop very slowly over the course of millions of years in the earth's crust, even though they are produced by forces that are not very great.

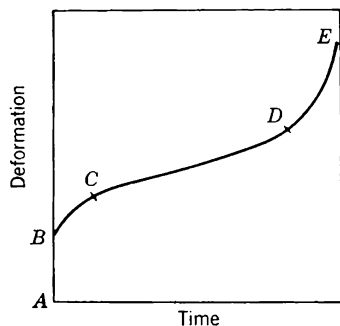


FIG. 162. Ideal curve of creep.



FIG. 163. Diagram of slip planes in extension.

Plastic deformation is directly connected with the displacement of particles in shear. In small-scale deformations, the shear displacement of the particles is approximately in the directions of the greatest tangential stresses. In these directions, slip planes, along which the particles are displaced relative to each other, are formed within the body. In plastically deformed specimens of metal, these slip planes, which in most cases coincide with the planes of the maximal tangential stresses, may be seen as the so-called Lüders lines. Similar lines can be seen on the faces of deformed specimens and of various other plastic materials. It has been established, for instance, that the plastic extension of a single-crystal wire takes place by a relative slippage of very fine plates along planes inclined at an angle of 45° to the axis of the tensile stress (Fig. 163).

The existence of connected planes of maximal tangential stresses produces a more complicated picture of slip within a body in tension and compression. Figure 164 shows a diagram of the slip planes in a

body subjected to compression. The slip planes divide the body into a multitude of tiny flat prisms, of which some are pressed outward to the sides and others wedged in between them. If the body undergoes shear, the slip takes place as shown in Fig. 165.

Theoretically, the slip planes should be formed at distances of one molecule's thickness from each other. In actual fact, these distances are always considerably greater and can scarcely be less than 0.0004 mm. They differ in different cases, depending on the material and the external conditions of the deformation. When the plasticity of the body increases (through a rise in the temperature or ambient pressure, a decrease in the rate of deformation, etc.—see below), the slip planes are closer together; conversely, the distances between them increase as the plasticity decreases (as, for example, through an increase in the rate of deformation).

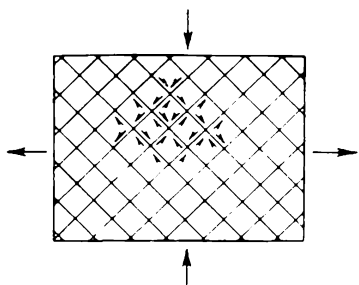


FIG. 164. Diagram of slip planes in compression of a plastic body.

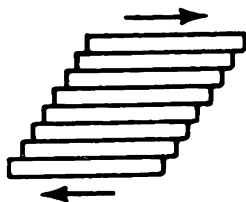


FIG. 165. Diagram of slip planes in shear.

Rocks that are composed of polycrystals, and particularly those whose structure contains discrete grains, display mechanisms of plastic deformation with specific peculiarities that have as yet been little studied. Rocks made up of grains usually contain "weakened" surfaces, along which the adherence of the particles is relatively small. In the deformation of such a rock, the slip planes that appear in it are not entirely new. These planes, as a rule, follow those of the weakened surfaces whose attitudes more or less coincide with the directions of the shear stresses.

Such weakened surfaces may be formed by the boundaries between the individual particles of the rock. In this case the slip does not penetrate into the grains, and the latter are not deformed. The deformation of the rock is accomplished almost entirely by means of the relative displacement of the grains; as they move, the grains rotate and adopt the positions most conducive to slip. Rounded grains continue to rotate as long as the movement takes place, whereas elongated or flattened grains that fall into the slip planes are rotated until they assume posi-

tions parallel to the movement. As a result of slip between the grains, the rock acquires an oriented texture parallel to the slip planes—that is, to the greatest shear (Belousov et al., 1951).

Slip in plastic deformation also takes place along the bedding planes and the contact surfaces between different geologic bodies, wherever these surfaces are more or less parallel to the direction of the greatest shear. Moreover, one may note a certain discrepancy between these planes and the greatest shear stresses, since it requires less energy to depart somewhat from these exact directions and to use the already existing weakened surfaces than to form completely new slip planes.

Movements between the grains are possible in relatively unconsolidated rocks, where the bonds between the particles are comparatively weak. Under such conditions, movement between the grains will not lead to rupture of the rock, because the deformation takes place under considerable ambient pressure and often because the grains are held

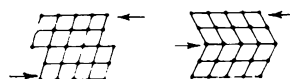


FIG. 166. Diagram showing the displacement of particles in translation (left) and twinning (right).

together by capillary water. In consolidated rocks, in which the grains are held together more tightly, it is frequently easier for the slip to take place not along the surfaces between the grains but within the grains themselves. Plastic deformation thus takes place by means of *intragranular slip*.

Within grains composed of crystalline material, the rows of points in the crystal lattice are the locus of the slip. If the slip involves relatively large surfaces within the crystal and cuts through all or a considerable part of it, it results in *translation*, or *twinning*, by producing a steplike displacement of the different parts of the crystal (Fig. 166). Slip within the grains has a different character if it takes place along a multitude of separate small surfaces differently oriented and not directly connected to each other. In this case, there is no slip of parts of the crystal extending throughout its entire volume, and the deformation does not take place along flat planes. This type of deformation frequently leads to recrystallization. Recrystallization is typically observed, for example, under such conditions in rocks subjected to axial pressure. The rock is crushed in the direction of its compression, whereas in the directions perpendicular to it the rock is expanded. In this process, the individual crystals change their shapes in the same manner, reducing their dimensions in one direction and increasing them in others.

Each of these two basic methods of deformation of rocks—intergranular and intragranular slip—may be manifested in any rock, depending on the conditions of the deformation. Intergranular slip evidently corresponds in general to lower temperatures, lower hydrostatic pressure, and more rapid deformation, that is, to the complex of conditions that as a whole is associated with relatively lower plasticity in the rock.

Intragranular slip, on the other hand, takes place under the conditions associated with greater plasticity in the rock: higher temperature, greater ambient pressure, and slower deformation.

Intergranular slip occurs along planes relatively far away from each other. In the case of slip within the grains, however, the distances between the slip planes are considerably less, so that the slip is observed to take place along a large number of surfaces very close to each other. Although it is intragranular, however, the slip may develop along planes whose average distance from each other varies greatly, and thus it may be more or less disseminated. More disseminated slip always corresponds to conditions of greater plasticity, and as the plasticity decreases, the slip becomes more and more concentrated, until finally it becomes intergranular.

It has been shown that in slip between the grains, the grains are rotated and they become oriented parallel to the shear surfaces. Here it is this rotation and displacement of the grains that determines the deformation of the rock as a whole, since each grain as a separate unit does not undergo deformation. In slip within the grains themselves, each individual grain is deformed, the changes in its shape repeating the deformation of the entire rock on a smaller scale. An oriented texture is formed in this case as well, since all the grains, being subordinated to a single overall deformation, are shortened and elongated in the same directions. The shortening is in the direction of the greatest compression, and the elongation is in the directions perpendicular to it—the directions of the greatest tension. The resulting oriented texture in this case differs in principle from that which appears in intergranular movement: in the latter case, the orientation of the grains comes about through their rotation and corresponds to the directions of the greatest shear, whereas in intragranular slip the orientation arises from the deformation of the grains themselves, so that their short and long axes are in the directions of the principal normal stresses, if the grains are to any degree isotropic.

Let us return to the graph showing strain as a function of stress (Fig. 161). The rise of the curve beyond the elastic limit indicates the occurrence of the phenomenon known as *strain hardening*. This is frequently observed in solid bodies, especially metals, and consists in an increase in the body's resistance to plastic deformation simultaneous with an increase in the plastic deformation itself. Thus it is believed that two processes are taking place at the same time: strain hardening, which increases the body's resistance to plastic deformation, and relaxation, which opposes this increase in the resistance. Since not much is yet known about the physical mechanism of these two processes, it will not be discussed here. The combination of these two opposing processes determines the nature of the deformation curve. Ordinarily, relaxation is

of great importance under conditions of slow deformation and high temperature, whereas strain hardening is more characteristic of rapid deformation and low temperatures. The curves shown below (Figs. 170 to 174) testify to the presence of this phenomenon in rocks and minerals, since the stress-strain relations indicate an increase in the stress as the deformation develops beyond the elastic limit. It is suggested, however, that this strain hardening is more readily manifested and has its origin not in the plastic deformation itself but in the hydrostatic pressure under which these experiments were conducted. This view agrees with the fact that, in contrast to metals, rocks deformed under high ambient pressure thereafter show less strength under normal conditions (Birch et al., 1949; Rostovtsev, 1935; Slesarev, 1936, 1948; Shreyner, 1950).

However this may be, in all types of deformation and in the case of all solid bodies, whether metals or rocks, the curves in these graphs, after rising to some extent beyond the elastic limit, level out more or less rapidly and finally bend downward. This "return" of the curve is especially clear in tension, although it is also observed in compression and shear. The downward bend of the curve means that further deformation takes place under decreasing stress; hence, in order to suspend the deformation, one must remove part of the load. In this one may see a process of reverse strain hardening, or decrease in the body's resistance to deformation. Ultimately this leads, as will be seen below, to rupture of the body.

In tension, the decrease in the body's resistance to deformation coincides with a transverse local narrowing of the body undergoing tension: this is known as *necking*. Experiments have shown that initially the plastic extension takes place uniformly throughout the entire body but that after a definite time it is concentrated in a limited zone, in which the area of the cross section decreases much more rapidly and a neck is formed. It might be thought that the decrease in resistance is caused by the contraction in the body's section in necking, but it has been shown that this decrease in resistance is greater than it would be if it were due only to this cause (Hanson, 1939). Thus one is forced to assume either a change in the internal structure of the material or a change in the mechanism of deformation.

Completely uniform deformation takes place only under ideal conditions. Usually, because of irregularities in the body's shape, external friction, nonuniformity of the structure, irregularities in the temperature distribution, and other factors, the plastic deformation is nonuniform. In uniaxial compression, for example, owing to the friction that develops between the body and its supports, the deformation is less in the outer parts of the body, in contact with these supports, than in its interior. Observations show that the free sides of the body are also

less intensively deformed (Fig. 167). Because of the nonuniformity of the deformation, additional stresses arise in the body: its peripheral layer, as a result of the more intensive flow in the interior, is subjected to tension, whereas the internal parts, being withheld by the outer layer, undergo compression. Tension in the outer layers may cause cracks to form in them.

The nonuniformity of the strain is also the cause of the fact that, during the deformation, each point in the body is displaced in the direction of least resistance. Since resistance produces external friction at the body's supports, and since the friction is related to the shape of the body, its deformation depends not only on the stresses in it but also

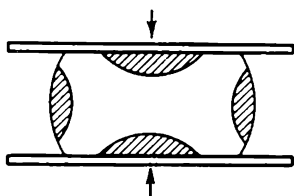


FIG. 167. Nonuniform deformation in the presence of friction on compressed surfaces; hatched areas are those undergoing smaller deformation (after S. I. Gubkin).

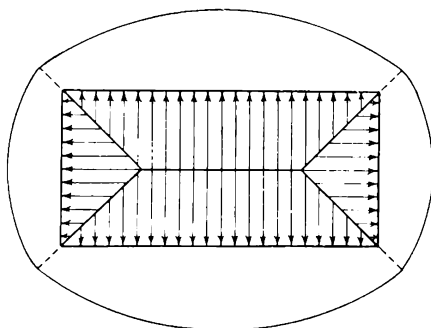


FIG. 168. Deformation of a parallelepiped caused by compression normal to the plane of the illustration; the inner contour shows the initial shape, the arrows show the direction of displacement of the particles, and the outer contour is the final shape.

on its shape. In uniaxial compression of a parallelepiped (Fig. 168), the particles encounter the least resistance in moving toward the nearest face. For this reason, the greatest elongation will be in the faces relatively closest to the body's axis; in other words, the smallest dimension of the parallelepiped will undergo the greatest increase. This process tends to force the body into a cylindrical shape with all points on the periphery at equal distances from the body's axis; thereafter, the deformation can take place at equal rates in all directions in the cross section.

A body's capability of withstanding plastic deformation without rupture determines its plasticity. The latter may be expressed, for example, in terms of the relative plastic elongation undergone by the body before it begins to rupture. Plastic properties are extremely varied. On the one hand, there are highly plastic bodies that can withstand great deforma-

tion without loss of their continuity (metals, for instance, may undergo plastic elongation of several hundred per cent). There are many bodies, on the other hand, that will undergo almost no plastic deformation; immediately after the elastic limit, or after only a very little plastic deformation, the body begins to rupture. Such bodies are called *brittle*.

The magnitude of the stress that produces rupture is called the *ultimate strength*. In the case of plastic bodies, there is a wide interval between the elastic limit and the ultimate strength, but in brittle bodies these two magnitudes are the same or almost the same.

It was formerly believed that each material has definite mechanical properties under all conditions, so that materials were classed as plastic and brittle, hard and pliable, and so forth. It is now well known that the same material under different conditions may have completely different properties: it may be brittle under some circumstances and more plastic under others. On the other hand, one may also select certain conditions under which the behavior of two materials whose properties are completely different under the usual circumstances (such as clay and metal) will be the same. Hence the mechanical properties are not individual and unchangeable peculiarities of the given material, but they are functions of both the material's structure and the external conditions.

The factors that affect the mechanical properties of materials are temperature, solvents and adsorbed liquids, ambient pressure, rate of deformation, and the stress state. The particular effects of each of these factors may be examined briefly below.

An increase in the temperature leads to an increase in the plasticity of the solid body. Heated metals, for example, are deformed much more easily than cold metals. Many materials that are brittle at low temperatures are capable of flow and of considerable irreversible changes in shape at high temperatures. This effect of the temperature is explained by the fact that as the thermal oscillations of the particles of the material are intensified, the bonds between them are weakened and their relative displacement is facilitated.

If the body is in contact with a solvent or an adsorbable liquid, this also increases its capability of undergoing plastic deformation. P. A. Rebinder's investigations in this field (1947) are particularly interesting. According to the results of his researches, the effect of the adsorbable substances is due to the presence of microscopic cracks in the surficial layers of the solid bodies. The liquid (such as water) is carried into these cavities by capillary forces and has a loosening effect, thereby softening the body. This involves a lowering of the elastic limit and an increase in the plasticity—the body's capability of undergoing irreversible changes in shape without losing its coherence. Adsorbed substances also affect the body's elastic properties, by producing a long-lasting elastic after-

effect and increasing the amount of elastic deformation, thus reinforcing the applied load in their loosening action. Sheets of mica, for instance, bend much more easily in moist air than in a dry atmosphere.

Figure 169 shows curves illustrating the change in the plastic properties of gypsum as a function of the action of the medium in which it is situated. Other experiments have shown that when marble is under a hydrostatic pressure of 10,000 atm under dry conditions, an additional directed load of 8,100 kg/cm² is required to produce a compression of 30 per cent, whereas in the presence of water saturated with carbon dioxide, a load of 1,500 kg/cm² is sufficient to produce the same deformation.

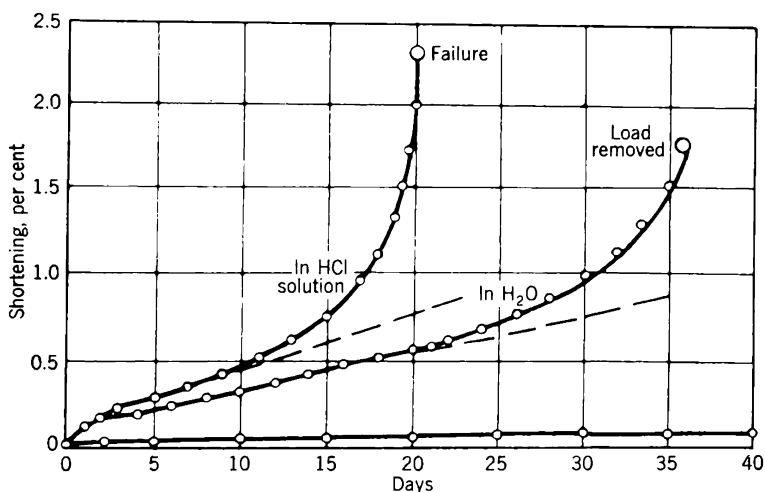


FIG. 169. Deformation of alabaster (in various solvents under a pressure of 205 kg/cm² at 24° temperature) (after Griggs).

Ambient pressure is an extremely important factor in the mechanism of plastic deformation and in the plastic properties of bodies, including rocks (Shreyner, 1950; Birch et al., 1949; Griggs, 1936, 1940a; Balsley, 1941). The effect of ambient pressure on plastic deformation is twofold. On the one hand, this pressure increases the body's resistance to plastic deformation, so that greater stress is required to achieve the same deformation under greater ambient pressure. On the other hand, the same factor strongly increases the elastic limit, the ultimate strength, and the body's plasticity. The increase in the ultimate strength means that the body can withstand considerably greater loads, without rupture, than under normal pressure. The increase in plasticity appears in the fact that the body is capable of a much higher degree of irreversible deformation without rupture than under smaller pressure. As a result of all this, a body that is brittle under normal conditions may become

plastic under high ambient pressure. It has been shown experimentally, for example, that a specimen of limestone, which under ordinary conditions undergoes only elastic deformation and ruptures as soon as the stresses reach the elastic limit, under an ambient pressure of 10,000

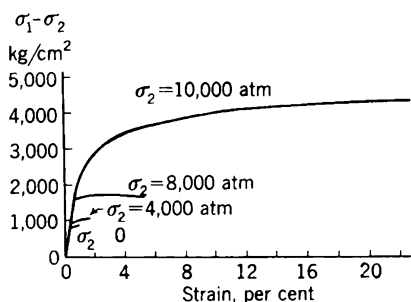


FIG. 170. Curve of the deformation of marble under ambient pressure (after Griggs).

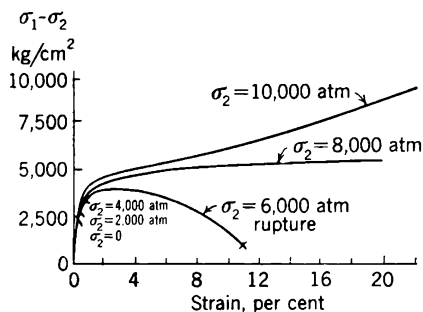


FIG. 171. Curve of the deformation of Solenhofen limestone under ambient pressure (after Griggs).

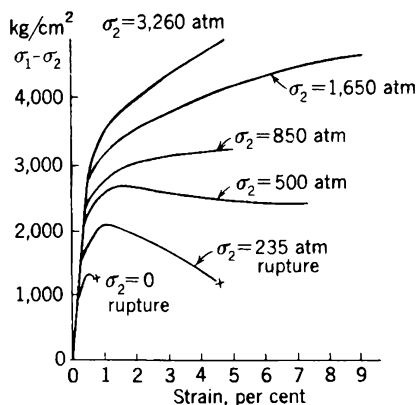


FIG. 172. Curve of the deformation of Carrara marble under ambient pressure (after T. Karman).

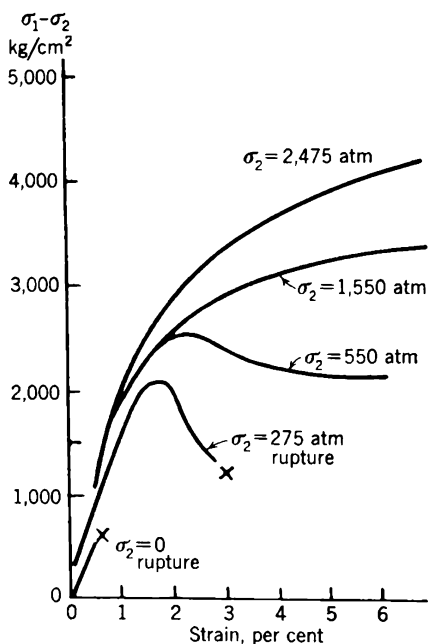


FIG. 173. Curve of the deformation of sandstone under ambient pressure (after T. Karman).

atm may be elongated by 50 per cent without rupture. Moreover, the limestone increases its ultimate strength from 2,600 kg/cm² at 2,600 atm to 13,000 kg/cm² at 10,000 atm. Under ambient pressures up to 4,000 atm, marble behaves as a brittle body; at a pressure of 10,000

atm, however, it may undergo plastic elongation of 25 per cent. At the same greater pressure, moreover, it withstands tensile stresses up to 4,800 kg/cm², sixteen times greater than the compressive stresses it will withstand under normal ambient pressure. The graphs illustrating these data do not require further comment (Figs. 170 to 174).

Under geologic conditions, the *rate of deformation* is obviously a basic factor affecting the plastic properties of rocks. In general, an increase in the rate of deformation increases the body's resistance to deformation and decreases its plasticity. Slow deformation, on the other hand, increases the plasticity. Therefore a body that is brittle when subjected to rapid mechanical action may undergo slow but considerable deformation under small but long-lasting stresses. The ultimate strength of the

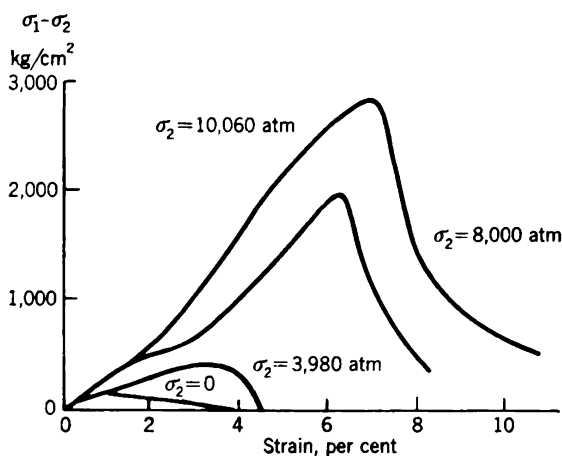


FIG. 174. Curves of the deformation of calcite under various ambient pressures (after Griggs).

material also changes: the stress required to produce rupture under rapid mechanical action is greater than under slow mechanical action.

All these laws are illustrated by the data on creep in gypsum given in Table 14.

Under rapid mechanical action in a dry state, gypsum has a strength of 520 kg/cm², and up to the time of rupture it undergoes a deformation of 0.35 per cent. Under the conditions shown in the table, any load from 125 to 300 kg/cm² produces rupture and in each particular case thus corresponds to the magnitude of the gypsum's ultimate strength. Since all the other factors are unchanged, the only cause of the change in the rock's strength can be the length of time. Actually, the table shows that the smaller the load, the slower and more long-lasting the deformation. The rate of deformation under a load of 300 kg/cm² exceeds the rate under a load of 520 kg/cm² by more than eighty times.

Table 14. Creep in Gypsum (After Griggs)

Load, kg/cm ²	Rate of deformation, millionths of the original dimension/ day	Amount of deformation before rupture, %	Duration of deformation before rupture, days	Equivalent viscosity, poises $\times 10^{16}$
300	2,000	0.89	2.45	0.42
250	440	1.39	13.54	1.60
205	219	2.30	48	2.64
181	100	3.00	133	5.11
165	77.0	3.6	285	6.04
150	66.5	>1.05	>110	6.36
125	24.5	3.8	308	14.4
103	7.0	>0.71	>520	41.0(?)

As the rate of plastic deformation decreases, its total magnitude increases. In the dry state, under a load of 520 kg/cm², gypsum very rapidly undergoes a deformation of 0.35 per cent and then ruptures. Under the conditions cited in Table 14, at a load of 300 kg/cm² and a relatively rapid rate, the deformation reaches 0.89 per cent; with a load of 125 kg/cm² and a very slow rate, it is as much as 3.8 per cent.

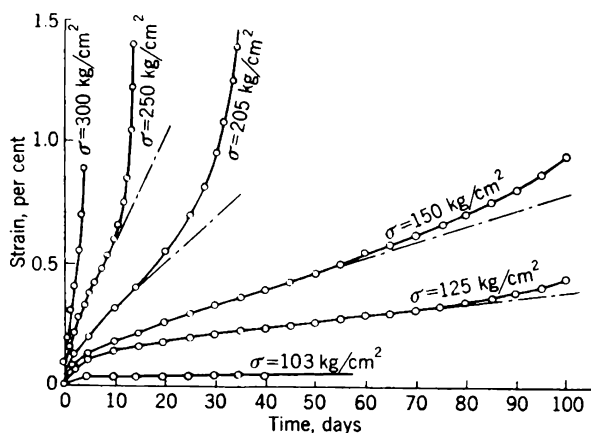


FIG. 175. Curves showing creep in gypsum under various loads, in an appropriately saturated solution, at 24° temperature (after Griggs).

Thus a small load produces a slow deformation, causing considerable change in the shape of the body before rupture begins. A large load, on the other hand, causes a rapid deformation, but the magnitude of the deformation before rupture turns out to be relatively small (Fig. 175).

The influence of the rate on the course of the deformation is closely associated with the speed of relaxation: a slow action exerted on the body

allows sufficient time for regrouping of the particles and fixing of the deformation, so that the latter may reach a considerable magnitude. When the action is rapid, the relaxation does not have time to go to completion and the accumulated stresses lead to rupture.

The last factor is the state of stress under which the plastic deformation takes place. S. I. Gubkin (1947), in his study of the properties of metals, showed that there is a definite succession in the different combinations of principal stresses in relation to their influence on plasticity and resistance to plastic deformation (Fig. 176). The most favorable stress states, in regard to their effect on the plasticity, are those in triaxial, biaxial, and uniaxial compression; the least favorable are the

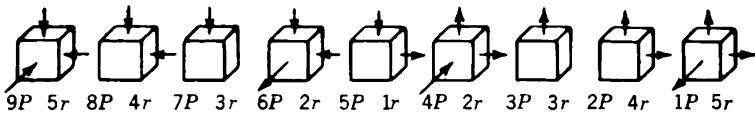


FIG. 176. Effect of stress states on plasticity (*after S. I. Gubkin*): P , plasticity, or amount of plastic deformation before rupture; r , resistance to plastic deformation; the figures indicate the relative magnitudes of the plasticity and the resistance to deformation.

stresses in extension. The effect of the stresses on plasticity is, in essence, the same as the effect of ambient pressure. It has been shown that any compression, even uniaxial, within the elastic limit must be accompanied by a decrease in the body's volume—that is, by ambient pressure, which increases the plasticity. Tension, on the contrary, increases the brittleness of the body and lowers its capability of undergoing irreversible deformation without rupture.

The body's resistance to deformation, according to Gubkin's data, is greatest when all the stresses are either compressive or tensile and smallest when some of the stresses are tensile and others compressive. Such systems of stresses with different signs are the most suitable for obtaining deformation of a specific magnitude with the least expenditure of force.

CHAPTER 19

FORMATION OF DISCONTINUOUS AND INTERMEDIATE FOLDS

ROLE OF A BURIED RELIEF IN THE FORMATION OF DISCONTINUOUS FOLDS

It has been suggested that some structures, which in our terminology belong to the category of discontinuous folds, are actually not tectonic formations but the surface reflections of an ancient buried relief. A. N. Mazarovich (1921), for example, has conjectured that some of the dislocations of the beds in the Russian plain are superposed or blanket structures covering protuberances in the ancient surface relief of the basement. Similar ideas have been suggested by American writers for flat discontinuous folds in the central states of the U.S.A. Another group of American geologists sees the reflection of the buried relief in the overlying strata not as a simple overlay but as a nonuniform compaction of the deposits over irregularities in the ancient topography, produced by the force of gravity.

If unconsolidated deposits settle on an uneven surface formed by solid, incompressible rocks, under shallow-water conditions the wave action of the water rapidly evens out the surface of the new deposits and keeps it permanently horizontal. At the same time the sediments are compacted by their own weight. The degree of compaction, according to experimental data and comparisons of the densities of rocks taken from drill cores obtained at various depths, is different for different rocks. According to American investigators, the maximal compaction is observed in clays: oozes immediately after deposition have a porosity of 70 to 90 per cent, whereas the porosity of clays long after deposition is always less than 30 per cent. The compaction of sands and limestones is almost nil (Hedberg, 1926).

It is suggested that if the clay series overlies an incompressible protuberance in the ancient buried relief, the compaction of the clay will be greater (in the sense of settling of the upper layers) away from this protuberance and less above it and on its slopes. As a result, the beds are bent into a domical shape over the protuberance; this bending, however, will be less pronounced in the upper layers than in the

deeper ones, so that the amplitude of the resulting fold will decrease upward in the section (Fig. 177).

The different capacities of rocks to become compacted under the force of gravity have led American researchers to think that in a complex section in which the lithology changes horizontally, nonuniform compaction of the deposits may lead to disturbances in the occurrence of the beds. Structural uplifts, for example, may be formed above lenses of sand, and depressions may appear wherever clay predominates in the section. Can we accept this hypothesis of the "passive" origin of discontinuous folds? To begin with, it must be stressed that this question can apply only to the most gentle folds. Dislocations of this type are common in the central states of North America; it is these to which the question refers. Other types of discontinuous folds distinguished by greater amplitudes and greater angles of inclination of the beds on the flanks, which are often broken by normal faults, quite clearly cannot be explained by this hypothesis. Moreover, gentle discontinuous folds are very closely associated with folds of other types, and in a number of cases (as on the Russian plain) combine with them to form a single structural complex. If it is also recalled that the discontinuous folds of different types belonging to a single complex are usually formed at the same time, it becomes extremely difficult to attribute to the most gentle folds an origin different from that of discontinuous folds of other types. If, however, this hypothesis can be applied only to very gentle folds, it cannot withstand criticism: when the sediments are subject to wave action, their surface rapidly becomes level.

The passive formation of such dislocations, according to the second variation of this hypothesis—that is, by the uneven compaction of sediments—may be observed only in a clay series. If one examines the sections through those parts of European Russia in which gentle discontinuous folds are widespread, one discovers that they differ little from the equivalent sections in the U.S.A. In the Volga-Ural region it is apparent that the rocks of the Devonian, Carboniferous, and Lower Permian systems that are involved in dislocations are represented almost entirely by limestones, dolomites, and gypsum; only in the upper parts of the section, in the Ufa, Kazanian, and Tatarian deposits, does one

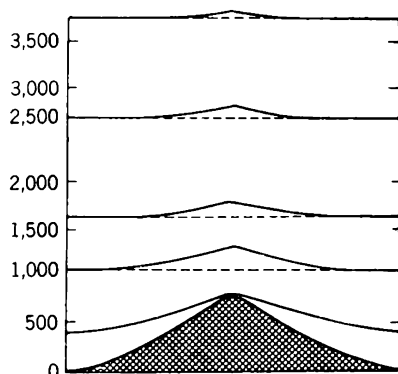


FIG. 177. Uneven compaction and settlement of deposits over a hard protuberance in the basement (after Nevin).

encounter sandy-argillaceous strata, and these are usually of small thickness. The lower, basal limestone part of the section is completely involved in the formation of the folds, so that these folds must have existed before the upper, sandy-argillaceous strata were formed.

In view of the fact that many of the folds show signs of repeated erosion, unconformable deposition, and repeated rejuvenation within a single locality, the conclusion becomes inescapable that the suggested mechanism of the formation of folds can have existed only in a small number of individual cases. It is, consequently, not a general mechanism applicable to all cases, but only one of the exceptions or derivations from the general rule that always introduce complications and variations into processes in nature. There can be no doubt that the formation of discontinuous folds as a whole must be considered not as a passive investment of an old buried relief by new sediments but as the result of active tectonic movement of the earth's crust.

DIRECTIONS OF THE FORCES FORMING DISCONTINUOUS FOLDS

After concluding that discontinuous folds, even in their weakest form, are produced by tectonic forces, it is logical to ask: what was the direction of these forces—horizontal or vertical? The best solution to this problem favors the second answer. It was said earlier, in Chap. 18, that discontinuous folds are characterized by the absence of linearity in the horizontal distribution of the folds, by the individual nature of the separate uplifts, by their localization, and by the sharp differences between synclines and anticlines. All these qualities are inconsistent with forces acting in the horizontal direction. Tangential forces applied to a given region from without will inevitably produce a regular linearity in the folded structure, which must thus show a similarity between the synclines and anticlines. These are all features characteristic of continuous, but not of discontinuous, folds.

The structural conditions that are manifested in areas of discontinuous folding indicate that such folding must have originated as a result of vertical, upward forces that uplifted each fold separately. The existence in such a structure of anticlines alone, with a complete absence of clearly formed synclines, is the best confirmation of this view. It is also confirmed by the frequent occurrence of flexures and vertical faults in the discontinuous structure; neither can be understood except as the result of radial movements caused by vertical forces.

SPECIAL FEATURES OF THE DEVELOPMENT OF DISCONTINUOUS FOLDS IN TIME

It has already been shown that one of the basic features of discontinuous folds is the gradual disappearance of folds of this type as

one moves upward in the section: their amplitude is smaller in the upper strata than in the deeper beds. This is clearly due to the fact that the thickness of the beds is smaller over the crests of these folds and increases elsewhere (see Fig. 32).

There are two explanations for this circumstance: according to the first, the thinning of the beds over the crests of domes is caused by their being stretched by uplift and bending; the second states that the same effect is produced by gradual development of the dome at the same time as the deposits are accumulated.

In certain cases the vertical crushing and horizontal stretching of the beds are actually a significant factor; this will be discussed later. This phenomenon alone, however, is far from sufficient to explain all the peculiarities observed in the structure of discontinuous folds. Very often, as we have seen, the change in thickness is accompanied by a change in facies. Coarser sediments are encountered on the crests of discontinuous folds than on their flanks: the shales in the flanks become sandstones in the direction of the crests, etc. It has also been shown that over the crests some beds not only decrease in thickness but wedge out completely, so that the later beds here overlie the earlier ones with a clear erosional unconformity. The combination of changes in thickness and facies and wedging out can be explained satisfactorily only by the hypothesis that such folds were uplifted gradually and simultaneously with the deposition of the sediments. The accumulation of deposits requires subsidence of the earth's crust, so that the entire process must be thought of as a gradual rising of the folds against the general background of crustal subsidence.

If the uplift of the fold lags behind the general subsidence, the crest of the discontinuous fold also subsides, but more slowly than the areas between the domical uplifts; consequently, the deposits are relatively thinner over the crest. If for any period of time the uplift of the fold is more rapid than the general subsidence, the fold undergoes absolute uplift and there is erosion of the sediments, which is reflected as a gap in the section.

The presence of coarser facies over the crests of discontinuous folds is explained by the fact that the surface of this part of the bottom, because of the constant relative uplift of the dome, is here somewhat elevated over the rest (forming a shallow-water area) and the sediments are deposited under conditions of more intensive wave action. The waves, in sorting the material, leave the coarser and heavier fraction of the sediments here and carry off the finer particles to the relatively deeper parts of the bottom—the spaces between the domical uplifts.

Discontinuous folds may be considered as analogous to intrageanticlines or central uplifts (although, of course, on a much smaller scale), since the gradual and long-lasting uplift in geanticlines has the same

effect on the distribution of thicknesses and facies as in the case of discontinuous folds. Later it will become possible to carry this analogy much further.

In most cases the uplift of discontinuous folds takes place slowly, over a long period, and at a variable rate. The development of a discontinuous fold may be traced by comparing the distribution of thicknesses and facies in the individual formations. A time of relative acceleration of the uplift will be marked by suites showing greater decreases in thickness toward the crest of the fold. In contrast, suites whose thickness does not change within the fold represent periods of temporary cessation of uplift. Erosional gaps in the section clearly correspond to especially intensive uplift of the fold, to a "pulse" in this process. According to data from V. Ye. Khain (1950c), who studied the domes of the Apsheron peninsula, for example, the changes in the thickness and facies of the deposits testify that the domes arose at the beginning of the Miocene, or even the end of the Oligocene, but that their most intensive growth was in the Pliocene and Quaternary. The history of these domes has been studied in especially great detail beginning with the time of accumulation of the Produktivnaya (oil-producing) series. It appears that the growth of the domes during this time was extremely uneven, being sharply accelerated at some times and retarded at others.

The salt domes of the Emba region grew with lack of uniformity during the Mesozoic and Cenozoic. In particular ages (such as the middle of the Early Aptian and the Early Albian), formations of unchanging thickness were deposited. There were times in which the uplift of the domes was suspended. In other ages (such as the Hauterivian), the domes grew with a noticeable decrease in the thickness of the formations over their crests. At still other times, finally, the growth was intensified to such a degree that stratigraphic gaps and unconformities appear over the crests (before the Valanginian, the Aptian, the Turonian, etc.).

Such changes in the rate of uplift of discontinuous folds have also been observed in the domes and arches of the Volga-Urals region. For example, the Polazna-Krasnokamsk swell (near the city of Molotov) underwent particularly rapid uplift at the end of the Late Carboniferous and beginning of the Permian, at the end of the Artinskian and beginning of the Kungurian ages, and so forth. In all these cases, the times of accelerated uplift of the discontinuous folds are apparently much shorter than those of slower development.

Observations show that such intensive pulses in the long process of uplift of discontinuous folds are common to all the folds located in large areas encompassing entire tectonic regions, although in different folds the individual pulses may be reflected to varying degrees.

It is extremely important to note that these pulses coincide in time with the phases of continuous fold formation manifested in the adjacent folded zones, and with the periods of intensified ascending oscillatory movements throughout the whole territory in which the folds are developed. Thus in the Apsheron peninsula and the Emba region, the most rapid growth of the domes is observed during the deposition of arenaceous sediments, indicating a relatively greater uplift of the whole region, whereas the periods in which argillaceous sediments accumulate indicate retardation of the growth of the folds. The greatest gaps in the accumulation of sediments on the crests of the domes coincide with the times of uplift over a considerable territory, during the course of the general oscillatory movements.

In view of this irregularity in the course of the growth of discontinuous folds, one may distinguish phases in their history characterized by more rapid uplift of the folds. These phases, however, are not nearly as short-lived as those in the formation of continuous folds (see below).

THE RELATIONSHIP BETWEEN FORMATION OF DISCONTINUOUS FOLDS AND OSCILLATORY MOVEMENTS OF THE EARTH'S CRUST

One form of the relationship between oscillatory crustal movements and the formation of discontinuous folds has been established above: discontinuous folds rise against the background of general subsidence of the earth's crust. This connection between the two tectonic processes may be clarified. Discontinuous folding is associated predominantly with platforms. The conditions of its occurrence may be contrasted with those under which continuous folds are formed, which are encountered almost exclusively in geosynclines. Discontinuous folding, however, also occurs in geosynclines. But a study of such cases eliminates the apparent contradiction. One case is that of the very end of the geosynclinal cycle, when the geosyncline has terminated its development and its subsidence has been replaced by general uplift. Others are encountered in those parts of the geosyncline in which, during the first stage of its development, there is little subsidence or even some uplift (the intrageanticlines). In the second stage of development, it has already been shown, such zones sink and ultimately become synclinoria. It is in these great synclinoria that the discontinuous folding takes place. In the third group of cases, the discontinuous folds within geosynclines are observed at the plunges of large anticlinoria—again in areas of less intensive tectonic activity. Discontinuous folds are widely developed in foredeeps.

An example of the first case is the Ural-Kara-Tau area (the southern margin of the Ufa plateau). This is a vertically uplifted block bordered

by flexures in the Lower Permian deposits and by steep faults. Its structure, as reflected in the Lower Permian rocks, is that of a box fold of the discontinuous type. Within this block, however, one may see continuous folds in the Lower and Middle Paleozoic rocks that have been unconformably cut through by the boundaries of the block. Thus uplift of the block took the form of the emergence of a box fold in an area occupied by earlier continuous folding.

Examples of the second case have been cited earlier; nevertheless, one may recall the Kura depression of the Caucasus with its domical discontinuous folding. The third case may be observed, for example, at the plunges of the main Caucasus range: the Apsheron and Taman peninsulas. There are numerous examples of foredeeps with discontinuous folding, such as the Cisuralian and Ciscarpathian foredeeps.

After the end of its development, the geosyncline is transformed into part of a platform; according to their regime of oscillatory movements, in the first stage of the cycle at least, the intrageanticlines resemble platforms and undergo no considerable subsidence. The foredeep, as far as its tectonic conditions are concerned, is a transitional zone between the geosyncline and the platform. All these observations point to the fact that conditions of little tectonic activity—small subsidence and moderate uplift—promote the development of discontinuous folds in geosynclines. On platforms, however, the discontinuous folding occurs in areas of relatively greater crustal subsidence—in the subgeosynclines.

The eastern Russian and Dnepr-Donets subgeosynclines are the areas of the most intensive discontinuous folding on the Russian plain. Such folding also occurs in the subgeosyncline of Poland and Germany; the Subhercynian, Thuringian, and northern German subgeosynclines; the Paris and Aquitaine basins, etc. On the North American platform, folding of this type is also predominantly encountered in areas of subsidence of the earth's crust: in the so-called geosynclines (actually, sub- or parageosynclines) of the Rocky Mountains, the Gulf of Mexico, the Illinois "basin," the Permian "basin" of the central states of the U.S.A., etc.

Regions such as the Voronezh subgeanticline and the great Baltic subgeanticline (the Baltic shield), the central plateau of France, the Armorican and the Rhine subgeanticlines, and others as a rule contain no typical discontinuous folds. Their structure is characterized rather by the predominant development of disjunctive dislocations, or faults.

Discontinuous folding, however, is not manifested to an equal extent in all subgeosynclines. One example of a subgeosyncline with weakly developed folding of this type is the Moscow basin; the reason for this difference between some subgeosynclines and others will become clear below.

Uplift of discontinuous folds usually begins some time after the

beginning of the subgeosynclinal subsidence in the given territory. The Transvolga region began its subsidence and became an area of sedimentary accumulation partly at the beginning of the second half of the Middle Devonian; discontinuous folds began to rise, wherever this can be determined by the distribution of thicknesses and facies, at different times, some of them in the Devonian and certain others only at the end of the Artinskian age (Early Permian), or even later. In the last case it is only beginning with Artinskian or later beds that one notes a decrease in thickness and a coarsening of the facies at the crests of the folds. The Subhercynian and Thuringian subgeosynclines began to subside at the beginning of the Triassic, whereas the folds began to be uplifted only at the end of the Jurassic period. The Paris basin began its subsidence in the Early Jurassic, but discontinuous folds appeared in it only at the end of the Cretaceous. The duration of the folds' uplift is evidently limited by the subsidence of the subgeosyncline: the domes continue to rise only as long as the general subsidence of the locality continues.

Stille (1912), in studying discontinuous folding in central Germany, found that the complexity of the folding increases from places with deposits of relatively small thickness to places in which the thicknesses are greater; in other words, the intensity of folding increases with the increase in the amplitude of the crustal subsidence. This rule is also mentioned by Bubnoff (1936), who noted that the more intensive uplift of the folds corresponds in time to the epochs of more rapid general subsidence of the earth's crust. He pointed out that Europe contains a progression of basins with subsidence of ever increasing amplitude, from the Moscow basin in the east through the Polish-German basin to the Subhercynian and Thuringian basins in the west. "Platform" folding also increases in intensity in the same direction, from the weak folding in the vicinity of Moscow to the gentle swells of the Polish-German basin and to the distinctly manifested folds of central Germany.

There are also certain indications of such relationships in the east of the Russian plain, where the great subsidence during the Late Paleozoic in the Cisuralian foredeep corresponded to a time of sharper discontinuous folding (in the Ishimbayevo district and elsewhere). On the lower reaches of the Kama River, where the thickness of the Paleozoic deposits is much less than in the Urals region, one finds that the isolated uplifts are much less distinct. One must not, however, attribute too much significance to this relationship. Bubnoff's examples are in error in one respect: the Moscow basin was in no way an area of little subsidence. On the contrary, the Moscow subgeosyncline underwent greater subsidence than, for example, the Subhercynian, but nevertheless it displays weak folding. Evidently the intensity of the discontinuous folding is affected not only by the amplitude of the

subsidence but also by the position of the subgeosyncline on the platform. In subgeosynclines located near the center of the platform (like the Moscow basin) only weak folding develops, whereas the subgeosynclines of the periphery, close to the geosynclinal zones, are characterized by much more intensive tectonic movement. But the intensity of the discontinuous folding depends not on the pressure from the geosyncline but mainly on the subgeosyncline's age. With the course of time, the dislocations gradually become weaker in the older subgeosyncline.

In general, the process as a whole may be represented as follows: subgeosynclines are formed as a result of differential vertical crustal movements within the area of the platform; a certain time after the beginning of subsidence in the subgeosynclines, areas of upward swelling appear, which then undergo long and gradual uplift against the background of continuing general subsidence; moreover, in places of greater subsidence, the discontinuous folds are, in general, uplifted more intensively. The same course of events can be traced in foredeeps.

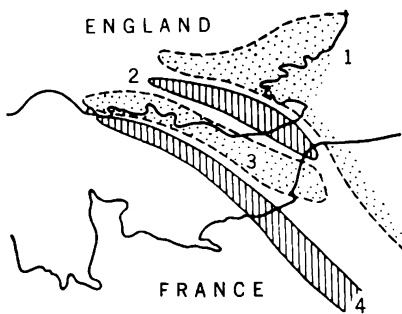


FIG. 178. Tectonic map of the English Channel region: (1) London basin; (2) Weald-Boulogne arch; (3) Hampshire basin; (4) Pays-de-Bré arch.

ment. The difference is that, in true geosynclines, the central uplifts ultimately lead to a complete replacement of subsidence by uplift and to a transformation of the geosyncline into its opposite: a mountain highland region. The growth of the discontinuous folds in the subgeosynclines does not go so far: subsidence does not cease to be the dominant movement; it is merely complicated by individual secondary uplifts. In addition, the central uplifts are areas of intensive continuous fold formation, which is not observed on discontinuous uplifts. This last difference, however, is partly contradicted by the formation of small concentric folds on certain discontinuous folds. Intrusives such as batholiths, finally, are formed in the central uplifts and do not appear in discontinuous folds. This difference is also not one of significance in principle, since igneous activity is far from unknown in discontinuous uplifts, although here it takes other forms.

The similarity between discontinuous folds and central uplifts is underlined by the fact that the former, like the latter, in certain cases

ing appear, which then undergo long and gradual uplift against the background of continuing general subsidence; moreover, in places of greater subsidence, the discontinuous folds are, in general, uplifted more intensively. The same course of events can be traced in foredeeps.

These relationships are analogous to the development of central uplifts within intrageosynclines. These, like the zones of secondary uplift, emerge from a background of overall subsidence and then develop through a long and gradual upward move-

arise as a result of inversion of the vertical movements of the earth's crust, inasmuch as they had already been "prepared" by the preceding crustal subsidence. The best example of such development of isolated folds is the Weald arch, which separates the London basin from the Hampshire basin and extends across the Channel into the region of Boulogne, in France, where it is known as the Boulogne arch (Fig. 178). This arch has been described by Lamplugh (1919). Drilling in its flanks has revealed that the anticlinal structure here occurs only in the surface strata (Tertiary and Upper Cretaceous). The Lower Cretaceous deposits are horizontal, and the Jurassic have an inverted synclinal structure. Thus the Upper Cretaceous anticline is superimposed on a syncline formed by the older beds (Fig. 179). The sedimentary layers form an enormous lens: the Jurassic and Lower Cretaceous deposits increase considerably in thickness (to 1,500 m) in the axial part of the arch and wedge out in the flanks, where the Lower Cretaceous directly overlies the Paleozoic. The Tertiary rocks, on the other hand, increase

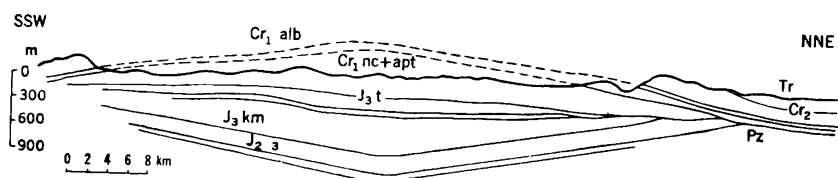


FIG. 179. Section through the Weald arch (after Lamplugh).

their thickness in the flanks. The Upper Cretaceous rocks retain approximately the same thickness throughout the section. Lamplugh comes to the only possible conclusion, that the sediments composing the arch, from the Triassic (not shown in the section) to the Upper Cretaceous, were deposited under conditions of crustal subsidence that was greatest beneath the axis of the structure. Toward the end of the Later Cretaceous the subsidence came to a halt; in the Tertiary it was replaced by the opposite tendency: uplift, which was again greatest along the axis of the arch. "The elevation of the accumulated deposits," wrote Lamplugh, "was greatest in the deepest part, where the deposits were thickest; and it is in this part that the uplift of the surface was formed." It is also characteristic that the stratigraphic gaps and the unconformities, which are very numerous at the margins of the Weald arch, disappear in the center of this fold.

The same relationships have been established in the Paris basin. According to Pruvost (1930), the anticlinal folds (swells) in this area correspond to the places of greatest sedimentary accumulation or subsidence of the crust—and the structural depressions are located over areas in which the Jurassic deposits are of small thickness and even

wedge out completely. Moreover, as in the preceding case, the Jurassic deposits form lenses under the arches, and the anticline on the surface is superimposed on a syncline in the deeper strata.

A number of similar instances have been encountered on the Russian platform in the Volga-Urals region. It has been established that in pre-Devonian times there was a basin of subsidence on the site of the present Oka-Tsna arch.

In the discussion of the morphology of discontinuous folds (Chap. 6), it was noted that the Volga-Urals region contains a number of cases in which the folds disappear with depth. This was attributed to the

increase in the thickness of the deeper strata toward the crest of the folds. These cases must quite clearly be classed with the examples considered immediately above, since here as well there was subsidence with a greater sedimentary accumulation than elsewhere on the sites of the folds before the uplift began. The succeeding uplift ("inversion") leveled out the downwarping in the lower strata and accordingly bent the higher beds upward. Wherever a more complex distribution of thicknesses is observed in discontinuous folds (see Fig. 33, right), one must assume a correspondingly more complex history of vertical movements of the earth's crust. It is interesting to note that in the case of

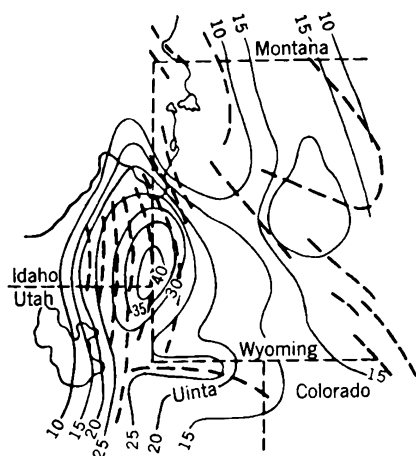


FIG. 180. Thicknesses of deposits and structure of part of the Rocky Mountains. Isopachs show the thicknesses of the Paleozoic and Mesozoic deposits in hundreds of feet; dotted lines are the axes of anticlines.

a domical uplift (the transitional form between discontinuous folds and anticlinoria) such as that of the Mangyshlak, it has been determined that the uplift developed in the Paleogene, on the site of a previous area of subsidence in the Mesozoic.

Where the uplift of the fold is "prepared" by preceding subsidence, its trend and disposition are linked to the disposition of the isopachs. An example of this is the Uinta dome in the Rocky Mountains. Figure 180 shows that this fold corresponds exactly to the area in which the Paleozoic and Mesozoic deposits are of considerably greater thickness than in the surrounding areas and thus form a structural depression. Consequently, here, as in the Weald, the local inversion in the formation of the box-type fold is quite evident. This relationship also explains the superficially peculiar east-west trend of the Uinta oval, which is at right angles to the predominant meridional trend of the folds in the

Rocky Mountains: this trend is entirely determined by the disposition of the area of previous subsidence.

Other examples have shown that when it is not possible to establish an individual preparation for each discontinuous fold by preceding sub-

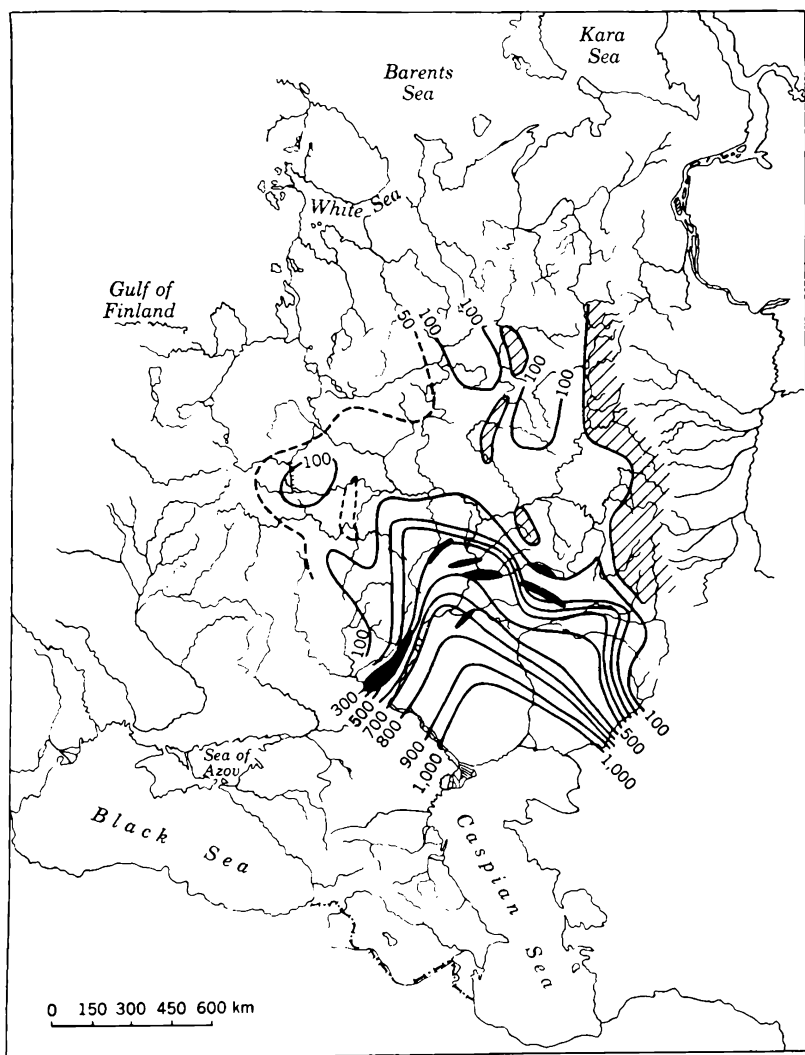


FIG. 181. Approximate distribution of total thicknesses of the Upper Devonian, Carboniferous, and Artinskian deposits (isopachs, in meters) and major discontinuous folds (solid black areas) on the Russian plain; the hatched area is the Urals uplift.

sidence but when the entire complex of folds has apparently been preceded by subsidence of the whole subgeosyncline, there is a general parallelism between the isopachs and the discontinuous folds. A fundamental example is the Volga-Urals region (Figs. 181 and 182). The first

of these structural sketch maps shows the isopachs of the Middle and Upper Paleozoic deposits and the disposition of the major discontinuous folds formed in the Paleozoic. The second shows the Mesozoic and Paleogene isopachs in conjunction with the folds that were uplifted dur-

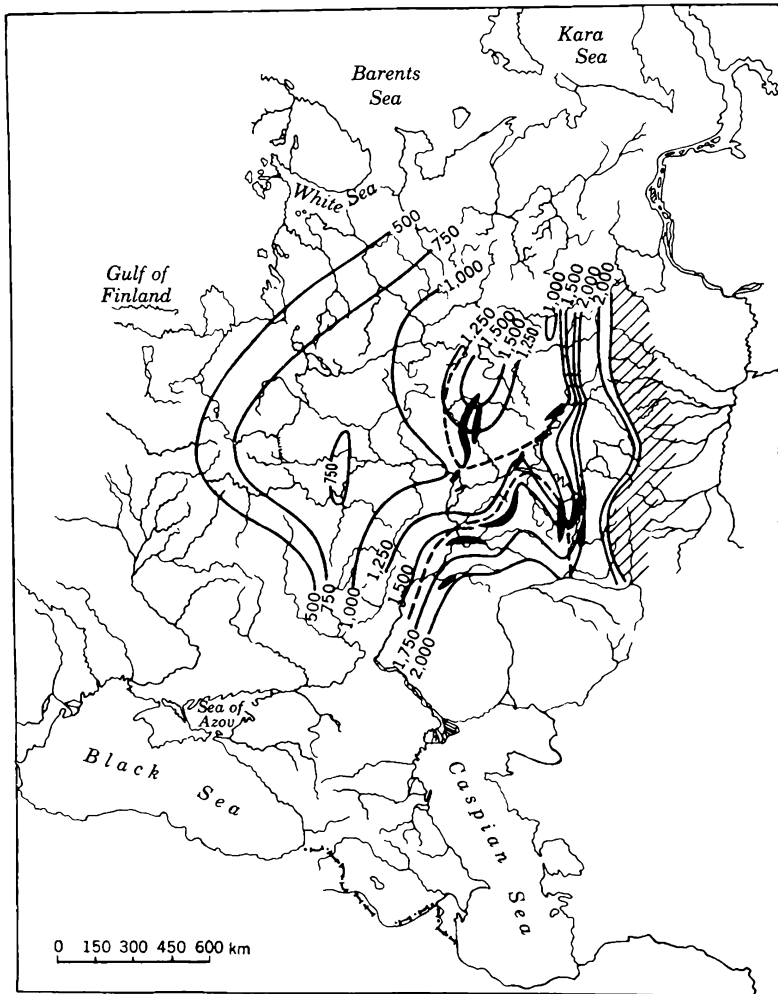


FIG. 182. Approximate distribution of total thicknesses of Mesozoic and Paleogene deposits (isopachs, in meters) and major discontinuous folds (solid black areas) on the Russian plain; hatched area is the Urals uplift.

ing the time represented by these systems. Both maps clearly reveal the parallelism between the folds and the isopachs.

This relationship between the disposition of the folds and the isopachs is of great interest. One may occasionally see a linear distribution

of these discontinuous folds, which at first glance resembles the linearity of continuous folding. Actually these are two superficially similar but essentially unlike phenomena. In continuous folding, the structure is determined by the regular and universal linearity of the folds and their mutually parallel disposition with a single predominant trend. In a discontinuous structure, the groups of folds may follow definite lines parallel to the isopachs; the presence of such lines is of great importance in practical work, for example in oil geology. But such lines are not necessarily an intrinsic feature, as in the case of continuous folds. In directly adjacent areas one frequently encounters folds disposed without any apparent order; the individual lines along which the folds are distributed may intersect, and they are not necessarily parallel; they bend, repeating the curvature of the isopachs; and such folds are not subordinated to any predominant trend extending over a large area.

The fundamental feature, as we have seen, is the fact that each discontinuous fold within a group of linear disposition is independent of the other folds in the group: different folds have different shapes, their axes are often oriented along directions entirely different from the trend of the group, and, finally, each fold has its own history of development. In other words, whereas in the case of continuous folding the structure is the result of a common general horizontal movement of the masses of rock composing the whole complex of folds, in discontinuous folding each fold is produced by local and independent manifestations of uplift, even if these individual manifestations may have a linear distribution over a certain area.

MECHANICAL TRANSFORMATIONS IN THE FORMATION OF DISCONTINUOUS FOLDS

Since discontinuous folds are formed by local uplifts of the crust, it may be concluded that the shapes of the folds, determining the fold type, depend on the intensity of the vertical forces, their distribution over the area involved, and the mechanical properties of that part of the earth's crust.

If the forces are intensive but confined to a limited area, they produce massive box-shaped uplifts with flexures in the flanks sharply delimiting the localities in which the forces of uplift are manifested. If comparatively weak forces are harmonically distributed over an area and they gradually increase in intensity toward the center and decrease toward the periphery, they will form gentle and indistinct discontinuous folds or swells. Fluctuations in the intensity of the uplift about a certain average magnitude produce complications of the basic fold in the form of domes of lesser orders; these subsidiary folds may be reflected with varying sharpness.

The mechanical state of the earth's crust determines its plasticity, and this, in turn, determines whether it will bend smoothly when the vertical forces are applied to it or be uplifted in separate blocks bordered by faults. These factors may differ in magnitude in each particular case, thus producing the great morphologic variety of discontinuous folds.

The reproduction of block folds in plastic laboratory models has shown that if a block is uplifted along faults in depth, folds of boxlike shape are formed only in the beds directly overlying this block. Higher up, the folds regularly become gentler upward, so that in the upper layers of the model the box folds are replaced by a gentle and broad archlike uplift of the beds. Thus the gradual upward leveling out of the beds in discontinuous folding may be to some degree due to mechanical causes and not alone to the long duration of the folds' formation.

Uplifting the beds in the earth's crust increases their surface area and thus subjects them to tension. Observations have shown that the mechanical properties of the strata enable them, within limits that differ greatly under different conditions, to be stretched by decreasing their thickness. This is indicated by the widespread development of discontinuous folds belonging to the group of closed folds, not broken by faults. The crushing and stretching of the beds require one to acknowledge that, under the conditions in the earth's crust, bedded rocks have considerable reserves of strength with which they resist the forces of uplift directed from below and react to them by plastic flow, combining extension and decrease in thickness. Thus one may understand the formation of the secondary concentric linear folds on the flanks of discontinuous folds that have been mentioned earlier. These folds have been observed on many of the discontinuous uplifts on the Russian platform. The following explanation is suggested: the upward forces that cause the uplift of the folds encounter strong resistance from the beds they attempt to bend and stretch, and under these conditions they give rise to great vertical stresses in the crust, which lead to the crumpling of certain formations. Since the vertical pressing of the beds is greatest over the center of the uplift, where the greatest forces are acting, and decreases in intensity toward the periphery, it becomes mechanically possible for a flow of material to take place from the area of greater to that of less pressure—from the crest of the fold to its margins. Encountering resistance from the parts of the beds that are lower on the flanks or outside the area of the dome and do not take part in the flow, the material of the beds must form secondary folds. The formation of such folds is not to be expected everywhere, but only where the amount of the pressure, its distribution, and the nature of the rocks favor its occurrence.

A typical feature of these folds that complicate the flanks of domes is their constant association with a limited group of beds and the fact that

they die out both upward and downward in the section. This shows that the crushing together and squeezing out take place only in certain (probably the most plastic) beds. On the other hand, it is not impossible that some folds of this nature may have been formed as a result of gravitational sliding of beds, or flowage of their materials, into places where the surface load was weakened by erosion (see below).

THE MIGRATION OF DISCONTINUOUS FOLDS

Earlier, in discussing the morphology of discontinuous folds (Chap. 6), mention was made of the lack of correspondence between the attitudes of different strata within a single fold. It was also shown that there are cases in which a single uplift divides with depth into two and that there is sometimes considerable displacement of the crest of the uplift. Such lack of correspondence was attributed to irregularity in the distribution of thicknesses in individual strata: areas of less and greater accumulation of sediments do not coincide vertically at different levels but are displaced laterally from one stratum to another.

Since the thickness of the deposits is determined mainly by oscillatory movements—by the distribution of less and greater subsidence of the crust—this lack of correspondence in the structure must be explained by the supposition that the distribution of the vertical crustal movements within the area of the dome or swell must have changed somewhat with the course of time. The changes involved migration of the center of maximal uplift, which did not remain in the same place but moved about within a certain area. They are also reflected in a more intensive or weaker differentiation of the movements, so that at certain times two domical peaks appeared where there had earlier been one, and vice versa. The history of all these changes may be uncovered only by a successive study of the distribution of thicknesses and facies in all the strata forming the given fold. The present structure of these strata is the net result of the combination of all the preceding movements.

The study of the migration and other changes in discontinuous folds with time is of great practical importance, since they may be the basis for predicting the structure of the deeper strata. Factual data, however, have only just begun to be amassed on this problem.

In the most gentle discontinuous folds, the ancient buried erosional relief is of some significance in the structural discrepancies between different strata. The sediments deposited on surfaces dissected by erosion to some extent repeat the latter's projections and depressions, and they acquire slopes that may not in every case coincide with the tectonic inclination of other beds. Since buried erosional depressions and projections are usually very small, this factor is concealed by major tectonic dislocations, including large discontinuous folds, but may be-

come noticeable in folds where the tectonic inclinations of the beds and the amplitude of their uplift are small.

DISCONTINUOUS BASINS

We have seen that a characteristic feature of discontinuous structures is the absence of synclines, which are replaced by residual depressions between the domes. In certain cases, however, within a region of discontinuous folds one may encounter structural forms that resemble

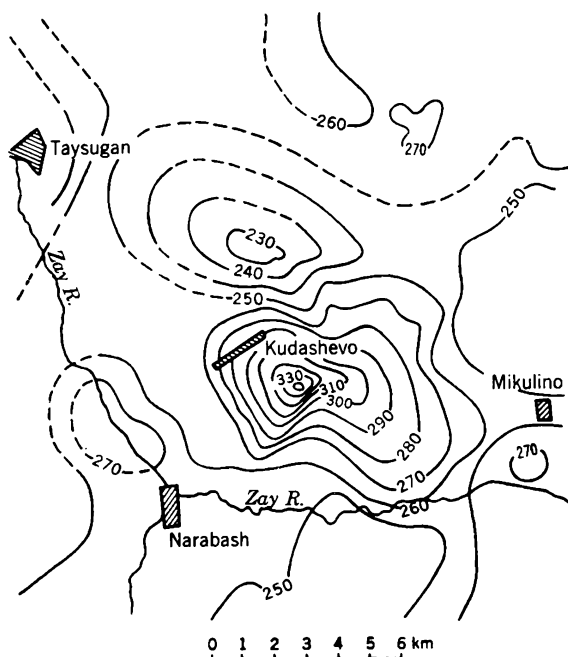


FIG. 183. Structural map of the Kudashevo district (Tatar A.S.S.R.): depression of the discontinuous type next to the dome. Structure contours show the *Spirifer* horizon.

sharply outlined basins: round basins or brachysynclines. As an example we may cite the almost completely circular basin that appears very clearly on structural maps of the area around Kudashevo, in the Volga-Urals region (Fig. 183). The narrow and elongated, very clearly delineated Ryazan-Kostroma structural basin should probably also be included among the structural forms of this type.

The lack of observations makes it impossible to discuss the nature of these structures with any thoroughness. There is no doubt that they are discontinuous or that they are isolated basins. It may also be noted that these basins are located next to positive discontinuous structures and

are, so to speak, their structural complements. Thus the Kudashevo basin lies at the foot of a dome of almost exactly the same size. The Ryazan-Kostroma basin accompanies and complements the Oka-Tsna swell. If this rule is confirmed, the discontinuous basins and brachysynclines may be seen as the smaller analogs of foredeeps and the attempt to draw an analogy between discontinuous folds and geosynclinal central uplifts will have still further basis.

CONDITIONS GOVERNING THE FORMATION OF DIAPIR DOMES

The fundamental aspect of the formation of diapir domes is the upward squeezing of plastic strata in the places where *piercement cores* are formed. Diapir domes are formed by a slow and gradual squeezing of salt or other plastic material through the surrounding rocks. This uplift, however, is punctuated by brief sharper upward thrusts, coinciding in time with the pulses in the growth of other surrounding discontinuous folds. The gradual uplift of diapir salt domes is also indicated by the regular changes in the thickness of the beds above the salt in the diapir domes (Table 15).

Table 15. *Thicknesses of the Deposits Overlying the Salt in Certain Diapir Domes of the Gulf of Mexico Region (After Yu. A. Kosygin)*

Age		Normal thickness away from dome, m	Average thickness over crest, m	Thinning over crest, per cent	Height of dome at top of suite, m
Oligocene	Vicksburg.....	360	12	97	1,230
	Frio.....	540	90	83	750
	Middle Oligocene.....	240	126	47	636
Miocene	Catahoula.....	15	...	100	600
Miocene and Pliocene	Oakville	990	342	65	37
	Lagarto.....				
	Goliad				
Pleistocene	Lissy and Beaumont.....	480	480	0	0

As shown in Table 15, the decreases in the thickness of all the strata toward the crests testify to the long uplift of the domes, combined with simultaneous subsidence of the locality as a whole and the accumulation of sediments. The fluctuations in this decrease in thickness at the crests indicate that the relationship between the intensive uplift of the domes and the rate of general subsidence of the area changed from age

to age. It must be said, however, that part of the decrease in thickness may be due to mechanical squeezing of the beds in the uplift of the salt domes.

The difference between diapir domes and other discontinuous folds lies in the sharply disharmonic relationships between the highly plastic rocks forming the core of the dome, on the one hand, and the underlying and overlying strata, on the other. The masses of very plastic rock force their way upward and pierce the overlying beds. During this piercement, the surrounding beds are bent upward and the crests are uplifted into domes, but these movements are always weaker than the movements taking place in the plastic series.

The interrelationships between the uplifting of diapir domes and oscillatory movements are no different from those generally typical of discontinuous folds, and they have been considered above: first, there is subsidence of the earth's crust for a certain time, followed by the continued growth, against the background of continuing overall subsidence, of the diapirs, punctuated by distinct periods of more rapid growth. In the southern Urals area subsidence began in the Middle Carboniferous, and the diapir domes began to rise in Ufa times (at the end of the Early Permian) and continued their growth to the end of the Paleozoic, as long as the subsidence of the region continued. In the Ural-Emba region the subsidence of the Alpine cycle (disregarding the earlier subsidence in the Hercynian cycle) began in the Early Jurassic, but the salt, which is of Permian age, began to be squeezed into diapir cores no earlier than the Middle Cretaceous and continued its upward movement through the Tertiary. The Permian salt in northern Germany began to rise in the Late Jurassic and continued through the Tertiary, in a region where subsidence began in the Triassic. In the Gulf Coast region of the U.S.A., the subsidence began no later than the beginning of the Cretaceous period, whereas the salt domes apparently rose during the Tertiary, while the general subsidence continued.

There is a view which holds that diapirs are associated exclusively with foredeeps. Although foredeeps are the predominant location for the development of diapirs, as a generalization this opinion is incorrect: the Ural-Emba province, northwest Germany, and the Gulf region are not foredeeps. Diapirs also appear in fairly intensively subsiding geosynclines, if they contain sufficiently plastic rocks.

The mechanism of the formation of diapir domes has attracted and continues to attract a great deal of attention among geologists, and a large amount of literature has been devoted to this problem. The special interest in diapirs is explained by the fact that many of them are the locations of oil deposits. A number of investigators have studied the mechanical properties of salt and gypsum in an attempt to determine the dynamic conditions under which these substances may undergo

such intense and concentrated plastic flow. Their investigations have been accompanied by experiments through which it has been established that salt, under unevenly distributed pressure, has an extraordinarily great plasticity and flows readily from points of greater to those of less pressure, deforming plastically on the way. Gypsum also has great plasticity under these circumstances.

The capability of these rocks to be deformed plastically is increased when they are in contact with solutions saturated with their own substance. Thus in regard to the behavior of salt, gypsum, and, of course, clay, under nonuniformly applied pressure, the problem of the formation of diapirs encounters no difficulties. The fundamental question here is the origin and distribution of the pressure. We shall not discuss certain views on the origin of diapirs that are now only of historical interest; the history of the various opinions on the formation of salt domes has been presented by V. A. Sel'skiy (1936). Among the older views of the origin of diapirs, one that still has some currency states that the cause of the uplift of diapirs lies in the increase in volume when anhydrite is altered to gypsum; this may be as much as 30 to 50 per cent (Kosygin, 1940). This view perhaps explains certain particularities of the formation of gypsum diapirs, but it is of no help in interpreting the phenomenon of diapirs as a whole. Not long ago the most commonly held view in Western Europe was that the squeezing of the salt is due to lateral tectonic forces of the same nature as those which produce linear folds. The particular form of the dislocation in diapirs, according to this theory, is caused by the great plasticity of the salt or other similar substance (Stille, 1917, 1925; van der Gracht, 1925). In addition, the adherents of this theory attempt to explain the round shape of diapirs by the intersection of folding in two directions (in Germany, "Hercynian" folding directed north-west and "Rhine" folding directed north-south).

The question of the possible participation of tangential pressure in the formation of discontinuous folds was discussed earlier and answered in the negative. Diapirs are a particularly sharp manifestation of discontinuous structure, and the theory of lateral horizontal pressure encounters special difficulties in being applied here: how is general lateral pressure to explain the formation, as observed in the Ural-Emba province and the Gulf Coast region, of circular vertical columns of salt in areas where the beds are elsewhere completely undisturbed? The idea of the crossing of tangential forces in two directions cannot be maintained, since from the standpoint of mechanics it is impossible to understand why two mutually perpendicular horizontal forces should cause the formation of such cylindrical vertical columns.

There is no doubt that the most correct current view is the one that sees the formation of diapirs as the result of local gravitational adjust-

ment of mass. This applies mainly to diapirs composed of salt. It is based on the fact that the specific weight of salt is less than that of the majority of rocks: hence if salt is overlain by heavier rocks, it will force its way upward through them and continue to rise until hydrostatic equilibrium has been established between the salt and the surrounding rocks. In other words, this view suggests that the salt is buoyed upward by the other, heavier rocks. The salt and the enclosing rocks behave as two very viscous liquids, one of which, the heavier, tends to move downward and the other, the lighter, to move upward. This opinion agrees with the presently accepted views of the mechanical properties of rocks and, in particular, of the role of creep in their deformation. It also accords with the specific weights of salt and other rocks: the specific weight of rock salt is 2.13, and the average weight of clastic rocks is 2.30.

This conception of the buoyancy of salt agrees with the shapes of the diapir cores (lenses, "stumps," columns, or "drops"). In this regard there are some interesting results of experiments on small-scale models of salt diapirs (Dobrin, 1941; Nettleton, 1934, 1943; Parker and McDowell, 1957; Lebedeva, 1956). Black bitumen with a small specific weight is poured on the bottom of a transparent vessel, and above this is poured a layer of heavier transparent sugar syrup. Some time after the layers have been poured into the vessel, depending on the viscosity of the materials in the model and the thickness of the layers, the bitumen begins to float upward. This buoyant material moving upward always takes the shape of a column or an inverted drop. Certain relationships have been established between the rate of growth of the diapir domes and their shape, on the one hand, and the difference between the specific weights of the two materials, their viscosities, and the thickness of the layers, on the other. If the thickness of the viscous fluid at the bottom (the "salt") is different at different points (because of the unevenness of the bottom), the diapirs are formed mainly in the zones of its greatest thickness; this agrees with what has been observed, under natural conditions, of the predominant association of salt stocks with depressions in the surface of the basement below the salt—the places in which the salt stratum is relatively thicker. Growth of the diapirs is also favored by an increase in the thickness of the upper stratum of heavier fluid. Observations under the conditions of nature have also shown that the growth of salt domes is facilitated by a great thickness of overlying rocks, but within certain limits: too great an increase in the thickness of the overlying rocks may, on the contrary, stop the growth of the diapir. This has been shown in experiments conducted by Parker and McDowell (1957), whose experimental conditions evidently most closely approximated nature. These workers, to represent the enclosing rocks, used not syrup, which has only viscosity, but sand, which

also possesses strength and a limit of plasticity. The existence of a plastic limit meant that when the thickness of the overlying layer of sand was increased beyond a certain limit, deformation of the sand was impossible and growth of the diapir ceased.

In her experiments N. B. Lebedeva reproduced the growth of a group of "diapirs" arising simultaneously and established certain relationships involving the distances between the diapirs, the viscosity of the materials used for the model, and the thickness of the beds (Lebedeva, 1956).

The experiment with bitumen and sugar syrup ends with all the bitumen floating on the surface of the syrup. Such a "floating" of salt on the earth's surface under natural conditions has almost never been observed; but one should not overlook the "salt glaciers" of Iran, which have flowed out upon the surface from the hidden cores of diapir domes.

Fissures in the overlying rocks facilitate the uplift of salt and other plastic rocks. In the Bashkir part of the Urals region, for instance, the diapir piercement cores have the shapes of rings or ovals and irregular rectangles (Fig. 184). These rectangles are several kilometers in cross section. At their surface the piercement cores are formed of gypsum and anhydrite; at greater depths there is salt. These rocks are of Kun-garian (Early Permian) age. Within each frame is a basin filled with the red-colored rocks of the Ufa suite, younger than the salt (dating from the end of the Early Permian and probably also the Late Permian). This unusual shape of the diapirs cores is most likely due to the correspondingly rectangular disposition of the tectonic faults in the series of rocks overlying the salt.

An interesting example of diapir folds is seen in western Provence (southwestern France). Here the Mesozoic deposits form a number of broad, gently sloping synclines separated by steep and narrow anticlines. In the cores of these anticlines, from beneath the limestones, marls, dolomites, and sand and clay deposits of the Jurassic, Cretaceous, and Paleogene, emerge the highly plastic rocks of the Upper Triassic series, represented by gypsum, clay, marl, and salt. This series forms typical piercement cores, and in places the Upper Triassic rocks have been squeezed to the surface as tongues forming small surface nappes up to 5 km long (Fig. 185). There can be no doubt that this entire structure has been formed by the squeezing of the light and plastic gypsum and salt series from beneath the broad synclines by the pressure of the overlying rocks.

Naturally, the squeezing of plastic rocks, apart from their specific weight, must be associated with an appropriate relief of the earth's surface: such squeezing must take place where the crust has been lightened by erosion—in depressions in the topography.

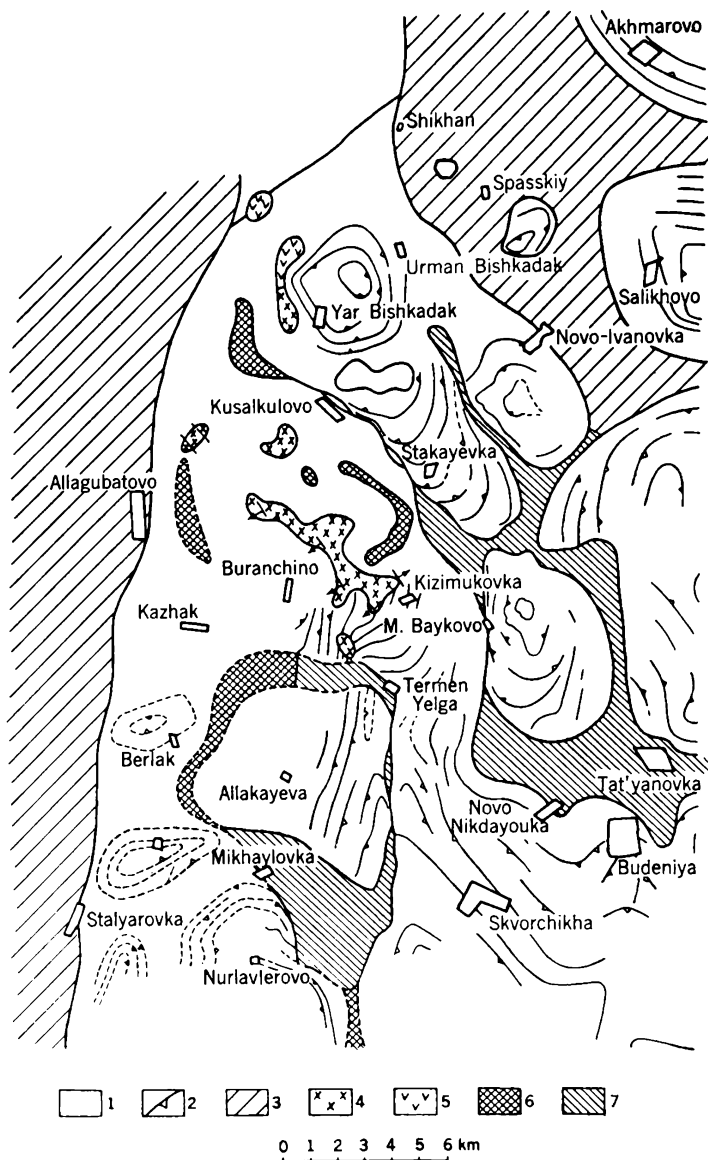


FIG. 184. Diagrammatic structural map of the Ishimbayevo district (after A. A. Bogdanov): (1) Ufa beds; (2) attitudes of Ufa beds; (3) areas of elevated Kungarian deposits within the region of gentle structures; (4) buried Artinskian massifs; (5) probable locations of massifs according to geophysical data; (6) covered diapirs; (7) outcrops of Kungarian rocks in diapirs.

The significance of the irregularity in the load of overlying rocks on the plastic series, in connection with the unevenness of the relief, can be seen from the example of the Jura Mountains (Fig. 186). Their structure consists of blocks trending northeastward, separated by vertical faults. In each block one may observe an overflow of plastic rocks of the Triassic (salt, gypsum, bedded marl) along the surface of the Paleozoic basement and beneath deposits of the Jurassic and Cretaceous (primarily limestones), toward the northern edge of the block, where the marls form piercement cores. This northward movement of plastic

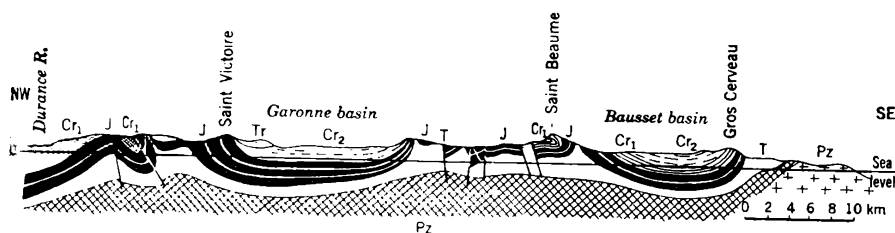


FIG. 185. Diagrammatic geologic section through western Provence (after Corroy and Denizot).

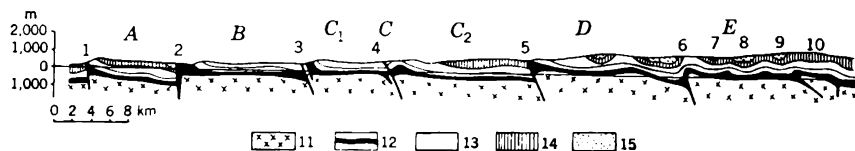


FIG. 186. General geologic section through the northern Jura Mountains (after Glangeaud): (A) block moved north from Ognon, (B) Besançon block, (C) Ornans block, divided into C_1 and C_2 by the Mamirolle fault, (D) Levier block, (E) Helvetic folds; (1) Saône fault zone, (2) Ognon fault, (3) Doubs faults, (4) Mamirolle fault, (5) Saline folds, (6) Laveron, (7) Planée syncline, (8) Saint Point synclinorium, (9) Metabief syncline, (10) Mont d'Or anticline, (11) crystalline basement complex, (12) Triassic, (13) Lias and Middle Jurassic, (14) Upper Jurassic, (15) Cretaceous and Tertiary.

material within each block is evidently due to an unevenly distributed load: because of the overall northward slope of the area, the southern part of each block is under a greater load than the northern, as may be seen from the illustration (Glangeaud, 1949-1950).

Certain peculiarities distinguish diapir folds whose cores are composed of plastic clays. Such folds are common in the eastern Crimea and in the Kerch, Taman, and Apsheron peninsulas. In plan, the clay diapirs are usually elongated brachyanticlines 5 to 20 km long and 2 to 5 km wide. In cross section, moving from the neighboring depressions through the flanks to the cores of the folds, the beds dip more and more steeply until they are finally vertical at the core itself. The cores are composed of highly crushed clay that has almost completely lost all

signs of bedding. Frequently one side of the core is bordered by a steep fault. Mud volcanos are closely associated with the clay diapirs, whose cross section is depicted in Fig. 187.

The clay diapir anticlines in these regions are usually divided by broad gentle depressions, from beneath which the plastic clay of the diapir cores has evidently been squeezed by the weight of the accumulated deposits in the basins. In the Kerch and Taman peninsulas,

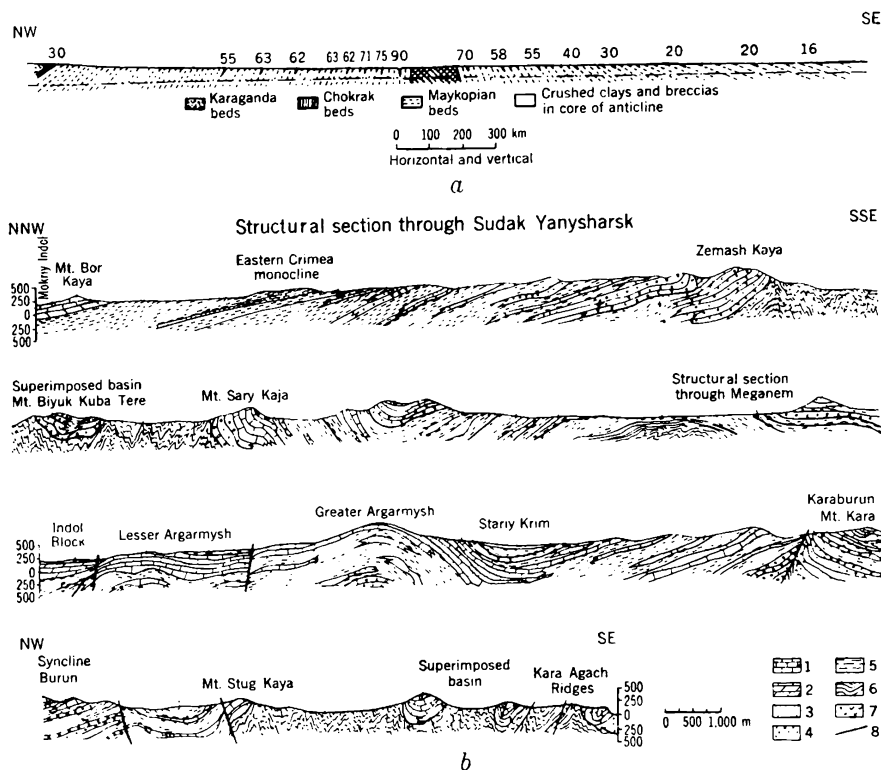


FIG. 187. Clay diapir folds: (a) Kerch peninsula (after A. D. Arkhangel'skiy); (b) eastern Crimea (after O. M. Filatov); (1) limestone, (2) marl, (3) sandstone, (4) conglomerate, (5) clay, (6) strongly crumpled shales of the Taurus suite, (7) volcanogenic rocks, (8) tectonic faults.

the plastic series that has been squeezed out is the Maykopian suite: clays of the Middle and Late Oligocene and Early Miocene. The cores of the larger diapirs in the eastern Crimea are composed of Lower Triassic and Lower and Middle Jurassic clay series.

A study of the history of clay diapirs in these regions shows that gentle swells were originally formed on their sites; these subsequently were slowly uplifted, at the same time as the depressions between them were being filled with ever thicker deposits. When the plastic

strata of the Maykopian series were exposed at the core of the fold by erosion, the clay began to be squeezed to the surface, with the formation of a diapir core (Fig. 188). The squeezing of the clays to the surface by the weight of the deposits accumulated in the adjacent depressions undoubtedly helped to increase the specific weight of these clays, by expelling the water and gas in them through the mud volcanos. In the

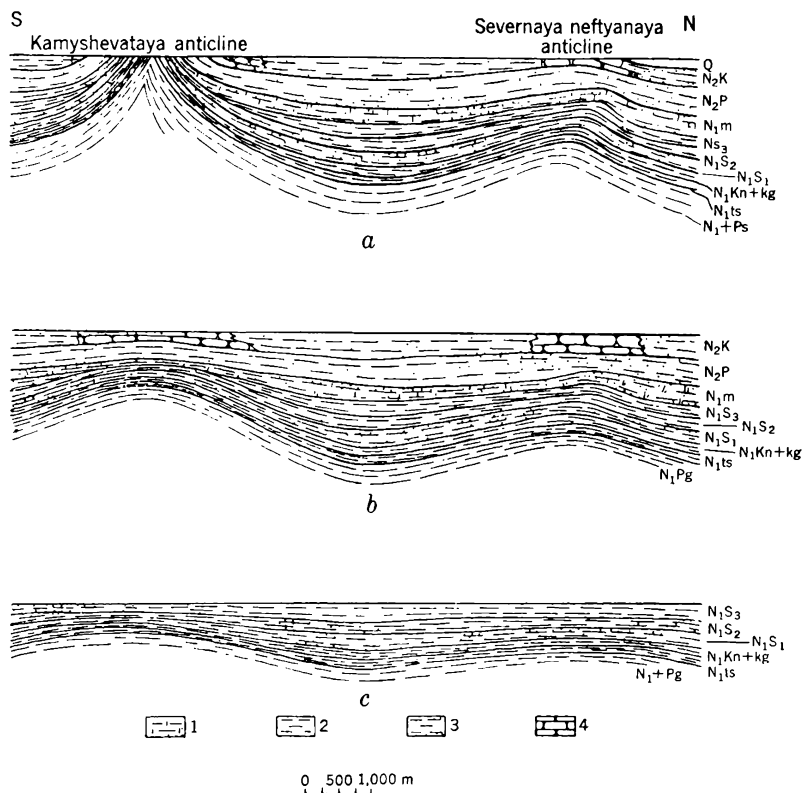


FIG. 188. The development of diapir folds on the Kerch peninsula (after N. B. Lebedeva): (a) present; (b) beginning of the Kuyal'nik age; (c) beginning of the Meotisian age; (1) sand, clay, and detrital limestone; (2) clays; (3) sands and clays; (4) sands.

Kerch peninsula this was facilitated by the relief, which was lower over the cores of the anticlines and higher over the flanks of the folds and the depressions. This last circumstance created an unevenly distributed load, which, in turn, aided the flow of the Maykopian clays toward the eroded cores of the anticlines (Belousoff, 1958; Lebedeva, 1958).

All the cases of formation of diapir folds are associated with such buoyancy, or squeezing out, resulting from the force of gravity. There

is evidently, however, still one more mechanism of formation of diapir folds. Figure 189 shows a section through the region of Iran in which the salt domes are developed (O'Brien, 1955). Such sections suggest that here the diapirs were produced not as a result of buoyancy but by the squeezing out of the plastic series from the crests of the domical uplifts toward the side during the growth of these uplifts, as the resistance of the overlying rocks exerted lateral pressure on them. The places in which the plastic series penetrated to the surface were probably predetermined by faults existing earlier in the overlying rocks.

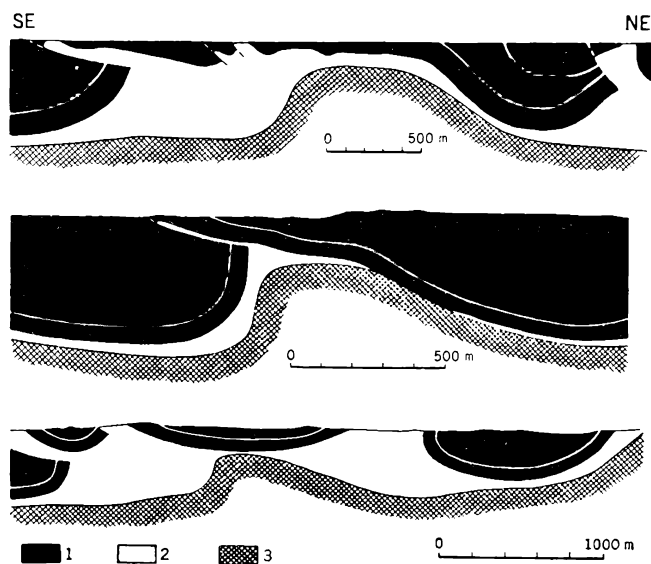


FIG. 189. Geologic sections through the salt-bearing region of southern Iran (*after O'Brien*): (1) upper hard series (Middle Miocene–Oligocene); (2) plastic series (Lower and Middle Miocene); (3) lower hard series (Lower Cambrian–Lower Miocene).

In conclusion, it may be mentioned that miniature diapir cores formed of coal have been observed. Coal has considerable plasticity under tectonic conditions. In the Chelyabinsk brown-coal basin, for instance, in places where the thick beds of coal have been completely or almost completely uncovered by surface workings, the coal is observed to be squeezed and considerably increased in thickness.

This discussion of the formation of diapir folds has concerned the general outer shapes of the piercement cores as a whole and has not touched upon their internal structure. Within salt and gypsum diapir cores, when there is bedding in the salt or gypsum, one may see a complex folded structure (Fig. 190). The stretched, reworked, and crushed folds of the salt and gypsum beds show that swelling of the plastic

series in the diapir core takes the form of an increase in the thickness of each layer separately and of a gathering of the squeezed mass into folds. The cores of the clay diapirs, as mentioned earlier, are almost structureless and composed of fine clay crushed into breccia.

Some mention must be made here of the peculiar complications in the shapes of the diapir folds observed in the Kerch peninsula and the eastern Crimea. In the Kerch area these have been called *impressed troughs*: at the center of the anticline's core, most often directly over its axis, is an almost completely round syncline composed of much

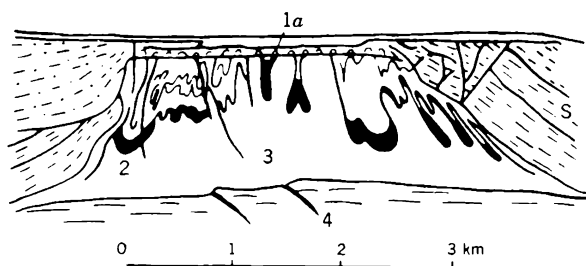


FIG. 190. Internal structure of a diapir salt dome: S, variegated sandstone, 1a, cap rock; 2 and 3, salt and anhydrite; 4, Hercynian basement (after H. Stille).

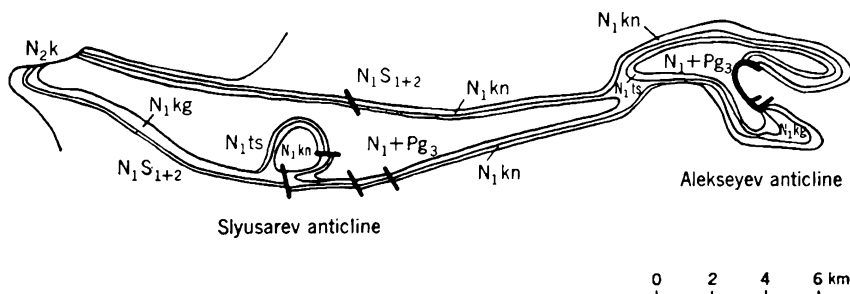


FIG. 191. Superimposed basins over diapir folds in the Kerch peninsula.

younger rocks (Sarmatian and younger deposits) lying unconformably and with a stratigraphic gap over the Maykopian clays of the core (Fig. 191). Whole groups of impressed or superimposed synclines of different shapes are observed over the crests of the diapir folds in the eastern Crimea (Fig. 192). In searching for an explanation, let us turn to Fig. 193. This shows the results of experiments with a model composed of layers of grease pressed upward at the center, as shown by the arrows. Under these conditions, the movement is primarily in the upper layers. As a result, in the compressed core one may observe a thrusting of its outer parts (flanks) toward the interior, which keeps moving upward.

Figure 194 shows one of the diapir anticlines of the Kerch peninsula

with a superimposed depression. Because of the similarity of this structure to the preceding illustration, it may be suggested that the more intensive squeezing of the Maykopian clays upward on the flanks of the folds may have led to the formation of secondary synclinal depressions along their axes and to the preservation of the higher strata overlying

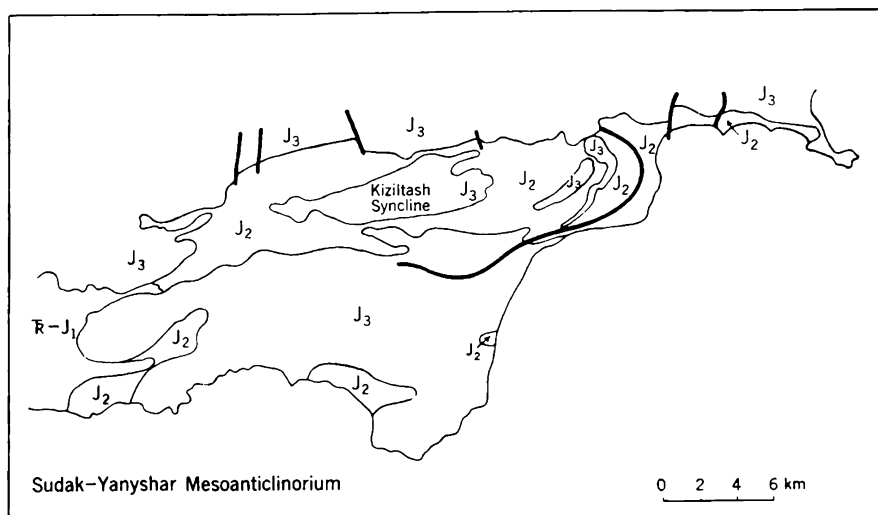


FIG. 192. Superimposed basin over a diapir fold in the eastern Crimea.

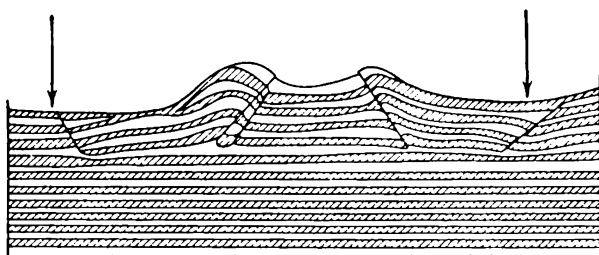


FIG. 193. Reproduction of pressure fold in a laboratory model; material is gun grease (after N. B. Lebedeva).

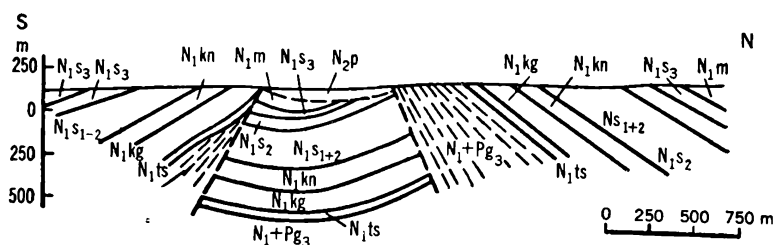


FIG. 194. Section through the Andreyev anticline on the Kerch peninsula.

the clays. Ultimately, blocks of these strata, being heavier than the Maykopian clays saturated with water and gas, sank into these depressions and gradually acquired the mechanically more economical shapes of round basins. The same process must be assumed to have occurred in the diapir folds of the eastern Crimea, where, however, the impressed basins did not assume circular shapes but formed elongated narrow synclines along the axial zones of the diapir folds.

CURRENT DEFORMATIONS OF THE EARTH'S SURFACE ASSOCIATED WITH THE GROWTH OF DISCONTINUOUS FOLDS

It has been shown that the growth of discontinuous folds is gradual and takes place over a long period of geologic time, punctuated by occasional intervals of rapid uplift. The effect of discontinuous folds on the distribution of facies and thicknesses (Chap. 6) indicates that during the time of their growth these folds were reflected by elevated areas on the surface. Discontinuous folds of different varieties—domes, swells, diapirs, etc. are growing in many regions at the present time and are reflected on the surface by local archlike swellings increasing with the course of time.

The present growth of discontinuous folds has been noted in the Pechora River basin, the Urals area, the Caspian lowland, the Apsheron peninsula, the Kura lowland, the Fergana Valley, the Tadzhik depression, and many other places (Nikolayev, 1949; Khain, 1950c; Lamakin, 1945, 1948; and others). Most often such active growth of discontinuous folds has been determined from geomorphologic indications. For example, the channel of a river flowing past the fold has been observed to move sideward, as if it were being displaced laterally by the growing uplift. Such is the origin of Samar Luka on the Volga, which is being displaced to the side by the growing Zhiguley uplift. If a river intersects a rising uplift or swell, there is a change in the structure of the valley: whereas beyond the area of the dome the river cuts through more or less thick sedimentary deposits, in the dome itself it cuts through the basement rocks. In this case, one may observe the deformation of river terraces, which acquire a domelike curvature.

Figure 195 shows a cross section through the Pechora River valley, past one of these present-day active uplifts. It is readily seen that the fluvioglacial and alluvial deposits are bent into a domical shape and that the older Quaternary layers are more strongly dislocated than the younger ones. The amplitude of the uplift varies: at the level of the fluvioglacial accumulations, it is 42 m, at the level of the ancient alluvial deposits it is 30 m, at the level of upper alluvial-erosional terrace it is 15 m, at the level of the lower alluvial-erosional terrace it is 9 m, and at the level of the old floodplain it is 9 m. At the core of the uplift

the river has exposed a moraine and Paleozoic deposits. This uplift is about 90 km in cross section: it is a large and very gentle swell, elongated from the southeast to the northwest.

The greater uplift of the older, Quaternary deposits signifies that the process of upwarping developed gradually during the entire Quaternary, producing greater dislocations in the older rocks. By rough calculations, the uplift is taking place at a rate of about 2 mm per year. Its present growth is reflected mainly in the narrowness of the valley and the greater curvature in the deviation of the channel than in the other regions and, as a result, in the more rapid flow of the water. This uplift evidently has a long history: to judge by the considerable dislocation in the Paleozoic rocks, it began to grow as early as the Paleozoic.

Similar things have been observed by L. N. Rozanov in river valleys of the Volga-Urals region, in places where the rivers intersect domes and swells being uplifted at present time. It has been established geo-



FIG. 195. Section along the axis of the Pechora River valley, with the levels of the fluvioglacial and fluvial deposits and erosions (after V. V. Lamakin): (I) level of fluvioglacial accumulations; (II) level of ancient fluvial accumulations; (III) level of river-cut terraces; (IV) old floodplain; (V) level of high floodplain; (VI) bottom of Pechora Valley at the river level. Deposits forming the valley: (1) fluvioglacial series of the Pechora Valley; (2) ancient alluvial series; (3) present-day alluvium; (4) moraine and Late Permian rocks; (5) elevation above Pechora River, in meters.

morphologically that the Vyatka swell, which began to form in the Paleozoic, is still rising at this time, as reflected in the relief of the locality, the intensity of erosional activity by the rivers, and so forth.

In the Kura lowland at the southern foot of the main Caucasus range, the Gerdymanchay River flowing down from the Caucasus, before emptying into the Kura River, branches into a number of stream channels that cut through a brachyanticline trending parallel to the latitude. This brachyanticline is clearly reflected in the relief. The branches of the Gerdymanchay River intersect this fold laterally, so that from a distance it resembles a loaf of bread cut into several pieces. It is quite clear that the cutting of the river channels occurred epigenetically, and simultaneously with the uplift of the anticline. This uplift continues to the present, as may be seen in the facts that the different branches of the river intersecting the fold cut through basement rocks, that the youngest Quaternary deposits are involved in the dislocation, and that this fold has quite evidently deformed the young leveled surface of the Kura lowland.

In certain cases, it has been possible to obtain quantitative data on

the movement of the earth's crust caused by the growth of discontinuous folds. The uplift of the domes in the Apsheron peninsula, for instance, has caused a noticeable deformation of the Baku water-supply line, built in 1917. Between 1923 and 1927 the duct located about 19 km from Nasosna Stanitsa was elevated about 1 m, and as a result it became necessary to build a new water-supply line. About 7 to 8 km from this point, however, the duct sank about 15 cm. I. I. Potapov has calculated that, over the course of about half a million years, certain Apsheron domes have been rising at the rate of about 1 mm per year (Khain, 1950*a*). The actual rate of growth during particular time intervals has been much greater, of course, since the uplift of the domes is uneven and at times has even given way to subsidence. In the Caspian depression, certain domes have been observed to rise at the rate of several millimeters per year.

The preceding data show that present-day folding movements, as reflected on the surface, are quite similar to oscillatory movements in their rate and general nature. They differ only in being localized and associated with individual folds of the discontinuous type. Some regularities in the disposition of presently rising discontinuous folds on the Russian platform have been noted by Yu. A. Meshcheryakov (1951*b*), who pointed out that the most active folds are near the margins of the platform.

The study of the present growth of discontinuous folds is of great practical interest, and a knowledge of such phenomena is necessary in the design and erection of large constructions: dams, canals, reservoirs, bridges, tunnels, water-supply lines, etc.

CONDITIONS GOVERNING THE FORMATION OF INTERMEDIATE FOLDS

Earlier in this book intermediate folds were divided into two kinds: box-shaped folds and ridge-shaped folds. There is scarcely any room for doubt that the folds of the box-shaped variety are the result of vertical movement of individual blocks of the earth's crust. In other words, the conditions under which they arise are similar to those leading to the formation of the domes, swells, and other varieties of discontinuous folding that are not associated with any diapir mechanism. On the other hand, there is some basis for thinking that ridge-shaped folding is in most cases connected with the diapiric squeezing of one plastic series or another.

The folding in Provence and the Jura Mountains, mentioned earlier in considering the conditions of the formation of diapir folds, is morphologically ridge-shaped, consisting of steep narrow anticlines and broad gentle synclines between them. The diapir folds of the Kerch

and Taman peninsulas are also ridge-like in their morphology. It is thus likely that in other cases as well the ridge-shaped folding occurs as a result of the flow of material in some plastic strata at depth from beneath the broad synclines into the cores of the anticlines, under the action of a load applied unevenly from above. In Provence and the Jura Mountains this mobile series was the Upper Triassic, and in the Kerch and Taman peninsulas it was the clays of the Maykopian stage. The displacement of the same Maykopian clays has evidently produced the ridge-shaped folds in northern Dagestan (see Fig. 34).

Such a movement of the material must have been facilitated by the gradual subsidence of the synclines, in which more and more deposits were being accumulated and thus continuously increasing in weight, and by the simultaneous uplift of the anticlines, where erosion was decreasing the thickness and weight of the rocks overlying the plastic series. If the plastic series is exposed by erosion, it may flow out on the surface, as has happened in Provence or the Kerch peninsula. But the displacement of material from under synclines into anticlinal cores can evidently take place even without these conditions, if the difference between the loads over the synclines and the anticlines is sufficiently great.

Hence the formation of ridge-shaped folds is promoted by the same differential vertical movements of the crust as those which directly determine the formation of box folds or discontinuous folds of nondiapiric character. But although in these cases the vertical movement of crustal blocks has caused the formation of these varieties of folding directly by elevating or lowering the beds, in the case of ridge-shaped folding the process is complicated by a horizontal flow of the material in the plastic series.

The dimensions of the blocks, the strikes of the boundaries between them, and the directions of the relative vertical movements determine the positions, sizes, and trends of the anticlines and synclines in ridge-shaped folding. This is the difference between the ordinary diapiric domes (salt stocks) and ridge-shaped folding: the shape of the former is predetermined simply by the buoyancy of the lighter material (the salt), which gathers itself into cylindrical columns and is drawn into each such column equally from all directions. Ridge-like anticlines, however, are elongated as a result of "directional" movement of blocks of the crust. But it should not be thought that a broad block is sinking beneath each broad syncline and a narrow block is rising beneath each narrow anticline; such an assumption is artificial. The structure of the Jura Mountains (Fig. 186) suggests that ridge-shaped anticlines may rise above faults that separate the individual great blocks if the movement of the blocks is not uniform and if one edge of the block rises higher than the other. Then, after erosion of the uplifted end of the

block, the plastic series is closer to the surface than under the sunken end of the block, where the load on it is greater. Thus there will be a flow of the plastic material toward the raised end of the block, with the formation of ridge-shaped anticlines along that edge, if, finally, the specific weight of the plastic material is lower or the surface relief is reversed (see above). Thus, in the case of ridge-shaped folding, a single syncline and a single anticline are connected in pairs within one deep-seated block of the earth's crust.

There are, however, folded deformations that belong morphologically to the category of ridge-shaped folds but whose origin is different from

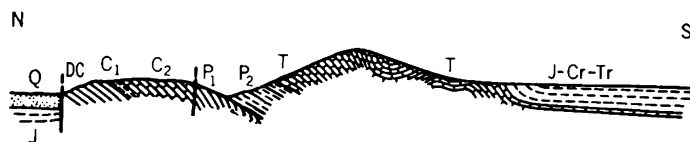


FIG. 196. Lutan uplift near Nanking.

that which has just been set forth. An example is shown in Fig. 196, where a number of almost parallel vertical faults divides the crust into blocks in the region of the Nanking hills in China. In the vicinity of the faults (usually on one side), the beds are sharply warped upward, forming ridge-shaped anticlines of which one side is cut by the fault. Since the oldest Paleozoic rocks are involved in the formation of the anticlines, they cannot be underlain by plastic sedimentary series whose flow may have formed these sharp anticlines. They should be seen rather as the result of some similar one-sided boxlike uplift. These may, however, be the deep roots of true ridge-shaped folds: if a plastic series had occurred above the visible section and had been covered by still younger rocks, there may have been a formation of ridge-shaped folds produced by subterranean flow of plastic material.

CHAPTER 20

MECHANICS OF CONTINUOUS FOLD FORMATION

Folding of the continuous type is a complex process of plastic deformation in an anisotropic medium. Its mechanics are still far from being sufficiently known. Only the most important aspects of this process will be dealt with here. For the time being, the discussion will be confined to processes taking place within either a single fold or a small group of folds, formed by compression of a certain segment of the earth's crust. The origin of this compression, the direction of the primary tectonic forces, and the mechanism of folding as a whole (meaning the conditions of formation of entire zones of folding) will be examined later.

COMPETENCE OF BEDS IN FOLDING

Willis has introduced the concept of *competent* and *incompetent beds*, which behave differently in the process of folding (Willis, 1932; Kirillova, 1949). According to Willis, competent beds are those rigid enough so that, when bent into a fold, they can lift the weight of the overlying rocks. Incompetent rocks are incapable of lifting such weight, and their role in folding is thus passive: they are supported by the competent beds. The latter, therefore, are the framework of a folded structure: they transmit the pressure and determine the shape and size of the fold. These concepts have gained general acceptance.

Nevertheless, a study of the internal structure of folded masses compels one to relinquish this concept and leads us to the conviction that the mechanics of folding are in reality quite different and that the terms "competent" and "incompetent" should not be used.

The first point to be considered is the secondary (mechanically caused) changes in the thickness of the individual beds within the fold. It has already been shown that a continuous folded structure is built up of "similar" folds, or folds with the same radius of curvature. In other words, in passing from one bed to another, the center of curvature moves correspondingly up or down by a distance equal to the thickness

of the bed. If, instead of layers of rock, the strata were composed of a stack of elastic sheets, they could have been bent into such "similar" folds only with the formation of empty spaces separating the layers in the crests and troughs of the synclines and anticlines. It is assumed that competent beds behave exactly thus: the distance between them increases in the synclinal and anticlinal crests and troughs and decreases in the flanks of the folds. No gaps are formed in the process, because the more plastic formations between the competent beds are squeezed from the flanks of the folds into the troughs and crests. Again, the more rigid (competent) beds are distinguished: in the process of bending, they press against the contiguous plastic beds.

Observations, however, show that the redistribution of material, manifested in fairly intensive folding by a decrease in the thickness of the beds on the flanks and an increase in the crests and troughs, affects not only the more plastic beds but also those that should be regarded as definitely competent. Varying degrees of redistribution of material are observed in heterogeneous masses, determined by variations in the plasticity of the individual beds and by the intensity of the compression (Kirillova, 1949). The most striking case is that of an entire series consisting of competent beds. There are no gaps between the beds in the crests and troughs, as should have been the case with the bent elastic sheets. All such voids are filled by the plastically redistributed material of the competent beds themselves. This redistribution can be seen in folded limestone beds with no interbedded plastic beds; in some folds the thickness of the individual beds in the crests and troughs of the folds is four times greater than on the flanks.

This common and well-known phenomenon flatly contradicts the entire concept of competence: a bed that thins out in one place and thickens in another cannot be regarded as a rigid layer supporting other layers. Such a bed is subjected to mechanical forces and reacts as a plastic body.

According to Willis, during the formation of an anticline, the bending beds lift the overlying formations. In that case, the maximum pressure should be concentrated at the apex of the fold, and if an internal deformation takes place in the competent bed, the redistribution of its material should have taken the form of a flow of material away from the crest and toward the flanks.

A mechanical decrease in the thickness of the beds at the apex of a fold is observed in discontinuous folds, as a result of localized vertical pressure directed upward. The opposite movement of the material in continuous folds points to a totally different mechanism of their formation.

It appears, moreover, that the shape of a fold is not necessarily determined by the mechanical properties of beds that would be con-

sidered competent in any given case. Rocks are folded into different types of folds, their shapes influenced by the plasticity of the formation as well as by the thickness of the beds. Plastic, thin-bedded rocks form sharper and steeper folds than those formed in rigid and massive rocks, all other conditions being equal. If, on the other hand, the section is composed of interbedded layers of different thickness and plasticity, the type of folding is determined, as shown by observations, by the dominant formations. Their dominance is manifested in their forcing other beds into shapes that are not characteristic of them under the given conditions.

From this standpoint, any heterogeneous section may be subdivided into *dominant* and *subordinate* beds. If, for instance, hard beds are interbedded with massive shales, the shape of

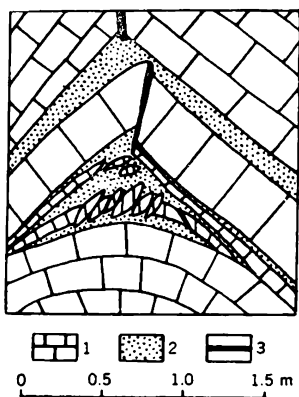


FIG. 197. Diagram of the structure of an anticlinal crest in Turonian limestones on the Rion River, Georgia (after I. V. Kirillova): (1) limestones; (2) marls; (3) line of fracture.

the folds will be determined by the shales and they will fracture. Rigid beds are incapable of stretching plastically on the flanks of a fold to the extent that softer beds are; for this reason a hard interbedded stratum on a flank is often found to be torn into pieces. Clearly, in all such cases, there is no question of any control by the hard beds: on the contrary, the soft, incompetent beds "impose their will" on the competent beds.

The leading role of the most mobile, plastic masses, which are more capable of redistribution, is clearly demonstrated by the various instances of disharmony. Figure 70 shows an example of a disharmonic fold. Here, at the base of the fold, lies a marl bed of greater plasticity and thickness than the contiguous beds. This accumulation of plastic material explains the formation of the fold. The effect

of plastic beds in the formation of disharmonic folds is well illustrated by another case observed in the same body of Turonian limestones in the valley of the Rion River (Fig. 197). The folded limestones form a gentle arch, above which an increase in the thickness of the marl bed is observed, varying from 0.05 m in the flanks to 0.15 m in the crest. The top of this marl layer is uneven, and the marl is sharply pressed into an overlying limestone bed, which in its loose portion is broken into lenses. The overlying, more massive limestone beds intercalated with marls have suffered a sharp fracture in their crests and are somewhat displaced vertically. Evidently, in this case the fracturing of the limestone was brought about by the pressure of the softer marl squeezed into the vault (Kirillova, 1949).

These instances, whose number could be multiplied, confirm that in the mechanism of folding, the active role belongs predominantly not to the hard but rather to the more plastic and mobile rocks.

ELEMENTS OF THE MECHANISM IN FORMATION OF CONTINUOUS FOLDS

The formation of continuous folds is a complex process combining several mechanical phenomena. The elements of this process are bending of the bed, redistribution of the material within the bed, and general compression of the entire series of folded rocks in the direction normal to the axial planes of the folds.

To clarify the significance of these elements of the folding mechanism, one must keep in mind the fact that a folding deformation encompasses the entire thickness of the folded mass. It is a fundamental error, when analyzing the conditions of folding, to consider them in respect to the bending of a single layer only, isolated and, as it were, suspended in space. All such approaches to the mechanism of folding are destined to fail, since no layer is isolated but rather is one of many that, in the process of deformation, mechanically interact closely with each other. The total deformed mass is always of tremendous thickness, in comparison with any individual bed or particular fold. Therefore, even if certain aspects of the folding mechanism are determined by some degree of independence of the individual beds, certain others are determined by their association with each other, their interaction, and their reaction to pressure as a single body.

The manifestations and the conditions governing the development of the several elements in the mechanism of folding may now be considered separately. The occurrence of bending of the beds in folding is self-evident. It is in this element of the folding mechanism that the role of the individual beds as more or less independent mechanical units is best manifested. Let us imagine an isotropic plastic body that, when compressed in the direction $c-c$, expands in the direction $a-a$ (Fig. 198).

There are no preferential directions in the body, so that its deformation is the same throughout. If it is further imagined that the body is stratified parallel to the plane $c-c$, regarding such "stratification" in the purely geometric sense without any mechanical significance, then each of these "strata," like the entire body, would contract along the axis $c-c$ and expand along axis $a-a$; that is, all the "strata," while remaining parallel to $c-c$, would increase in thickness. There is no reason for their being bent into folds.

The picture is altogether different when the stratification is not merely geometric, but mechanical as well, thus forming slippage sur-

faces and creating a generally heterogeneous body. In that case, a body compressed in the direction parallel to its stratification would not be flattened as a unit but would be bent and would form a number of folds. Such bending is accompanied by slippage of the layers past each other. This may be proved by a very simple experiment. If one takes a thick sheaf of papers or a deck of cards and fastens the sheets or cards to each other along their entire surfaces (by gluing them together) to prevent their sliding past each other, the resulting body could not be bent by hand. If, however, the sheets lie free, the stack is quite easily bent and the layers are observed to slide over each other. Figure 199 shows a stack of three layers (*a*) capable of sliding over each other. After the stack has been bent (*b*), the layers keep their original relative position only in the apex of the fold; some sliding of the layers over each

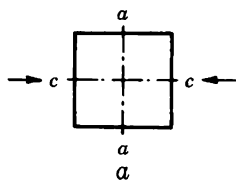
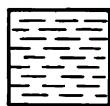
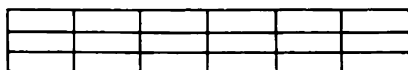


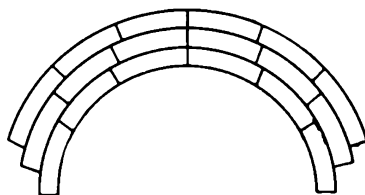
FIG. 198. Axes of deformation (*a*) and imagined stratification in an isotropic body (*b*).



b



a



b

FIG. 199. Diagram of the sliding of layers in bending: (*a*) original position; (*b*) final position.

other is inevitable in the flanks. As one can see by the marks, which had coincided before the bending, the displacement takes place in such a manner that each layer in the flanks slides over the one below it and toward the crest of the anticline.

The role of stratification in folding can also be demonstrated experimentally, on models built of plastic material. A sample was used, for instance, consisting of three thin wax layers and a thick clay layer above them. In longitudinal compression, the clay layer became thicker and formed a number of rough wrinkles, without forming any regular folds (Fig. 200). When the clay stratum was divided into four layers separated by wax paper to facilitate sliding, the new pack of clay layers formed well-defined folds (Fig. 201).

All other conditions being equal, the thinner the beds, the more easily they are bent. Therefore more complex and tighter folds, with steeper flexures, are observed in thin-bedded formations than in more massive formations.

Another factor, naturally, is the composition of the rocks and their mechanical properties: argillaceous beds form sharper and steeper folds than hard sandstone beds of the same thickness. Differences in rock composition and in the thickness of the folded beds are the essential factors causing disharmony in folding.

Sliding of the layers over each other in the process of folding produces certain additional phenomena, the most important of which is *shear cleavage*. The mechanics of this are the following: Because of friction, each layer, in sliding past both the overlying and the underlying layers, is acted upon by a couple (Fig. 202). One of the forces is applied to the top of the layer and acts in the direction of the rise; the other acts on the

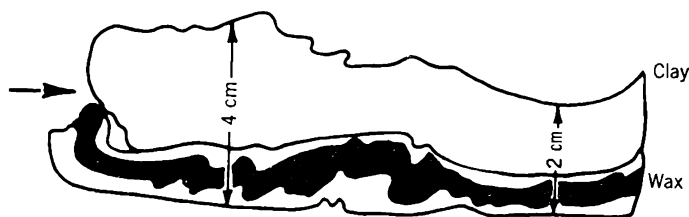


FIG. 200. Deformation of a thick clay layer.

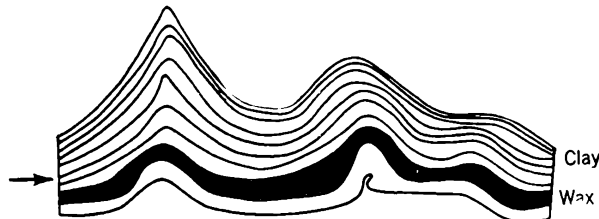


FIG. 201. Deformation of a stratified series of wax and clay layers.

bottom of the bed in the direction of the dip. The action of these two forces results in a shear deformation. In this process, two systems of parting surfaces are originated: one parallel to the layer and the other perpendicular to it. It is suggested that the shear cleavage, which is known to form a fanlike network in a folded structure, converging under the crests of anticlines, follows this second system of parting surfaces (Pek, 1940). Shear cleavage is manifested in each bed separately, independent of the other beds, and its frequency changes from bed to bed. Usually it takes the form of fractures perpendicular to the bed and located several centimeters apart.

Bending without an intraformational redistribution of the rock material within the bed is observed only in concentric folds, which, it is well known, are a rarity among folded structures. It has already been mentioned that continuous folding is usually manifested in "similar" folds.

whose formation is essentially characterized by, in addition to the bending and simultaneous with it, a plastic flow of the material from the flanks into the crests and troughs of the folds.

If bending is a manifestation of the mechanical independence of a bed, then the redistribution of the material within the bed must be considered as a result of the interaction of the layers in the course of folding of the entire series of layers. The beds are squeezed equally in the direction of the crests of the anticlines and the troughs of the synclines. In this process of redistribution, the bedding determines the intraformational nature of the flow. In beds of different rocks, the

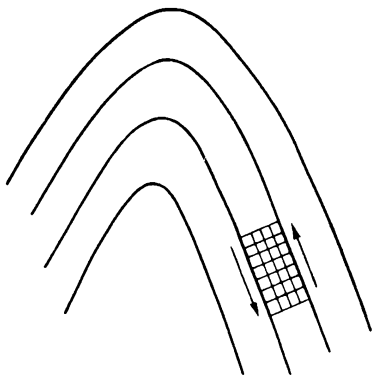


FIG. 202. Shearing in a layer in folding.

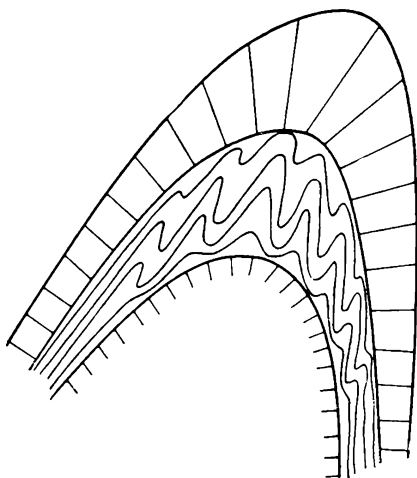


FIG. 203. Compression into small folds of plastic layers squeezed out of the flanks of a fold into its crest.

redistribution of their material varies in extent, depending on the plasticity. The degree of distortion of the thickness is naturally associated with the general intensity of the dislocation.

Because of the heterogeneous composition of the bed and because of the friction along the borders between the layers, the movement of the material on various levels within the bed, from the flanks toward the crests and troughs, takes place at different rates. Here a tendency is observed on the part of elongated mineral particles to orient themselves parallel to the bedding, or "along the flow." Thus an oriented structure within the beds, or schistosity parallel to the bedding, is formed as a by-product of intraformational flow; this is especially characteristic of ancient Precambrian formations. As a rule, such an oriented structure within the beds is better developed in the flanks than in the crests and troughs of the folds. In the latter, the orientation of the particles often cuts across the bedding, approaching the direction

of the principal flow cleavage and forming an inverted pattern converging toward the crests of the anticlines (see below). Frequently, the parallel texture of the crests and troughs is crumpled into minute folds (Fig. 203). The degree of flattening on the flanks of a fold, and the corresponding swelling at its crests and troughs, can be quantitatively determined when oolitic limestones are present in the dislocated series. In the course of flattening in the flanks, the oolites are also flattened, changing from their original spherical shape to ellipsoid. The compression of the ellipsoids—the ratio of their long and short axes—is the measure of the compression in the rock. For this purpose oriented slides are prepared, and the orientation and dimensions of both the long and short axes of a great number of deformed oolitic ellipsoids are determined. Thence the magnitude and direction of the compressive stresses are computed, and the amount of strain is established by a simple geometric construction (Cloos, 1947). Study of the deformation in oolites and in other inclusions whose original shape is known is of interest not only in determining the original thickness of the beds but also in comparing the intensity of the tectonic compression in various areas of a region of folding.

The last of the above-mentioned elements in the mechanism of folding is the general compression of the folded groups of beds. The presence of such compression is confirmed by the study of certain properties of flow cleavage parallel to the axial planes of the folds. This will therefore be called the *principal flow cleavage*. First, it is easily established that the principal flow cleavage is closely associated with the deformation of the rock—specifically, with its compression normal to the cleavage and corresponding elongation parallel to it.

This can be seen from the deformation of the various inclusions found in rocks. Organic remains found in rocks showing cleavage are always observed to be flattened along the axis normal to the cleavage and elongated parallel to it. Heim's description (1919) of belemnite specimens torn apart in the process of their elongation along the cleavage is a classic. Elongation parallel to cleavage is also suggested by the similar orientation of the *extension aureoles* found around pyrite grains and other solid inclusions (Fig. 204). In the compression and elongation of the pyrite cubes, their *shadows*—small cavities—are formed along the axis of greatest elongation. These cavities are filled with quartz, chalcedony, calcite, or some other mineral substance. Wettstein (1886) devoted a voluminous work to the deformation of fossil fish molds, confirming that their elongation in rocks with cleavage is parallel to the cleavage and their corresponding contraction normal to it. The same relationship

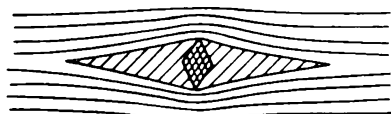


FIG. 204. Aureole of extension around a solid grain, in schist.

was established wherever it was possible to correlate the orientation of deformed oolites with the cleavage. Thus it is definitely established that a formation pervaded by principal flow cleavage has undergone compression normal to the cleavage and extension parallel to it.

Hence one must conclude that principal flow cleavage differs sharply from shear cleavage. The latter is related to shear and is oriented along the maximum shear stress. Flow cleavage is parallel to the long axis of deformation. Accordingly, the orientation of the grains that is produced by each type of cleavage has a different mechanism of formation. In shear cleavage, the flat grains are rotated into a position parallel to the direction of the relative slip of the cleavage surfaces. In principal flow cleavage, the orientation of the grains is a result of their flattening along the axis of compression and their elongation along the axis of

extension. In the first case, the grains merely change their positions and the deformation of the rock as a whole is achieved by means of their rotation; in the second case, the grains are deformed along with rock. Thus parting parallel to the texture of the rock is due partly to rotation of the grains and partly to their deformation.

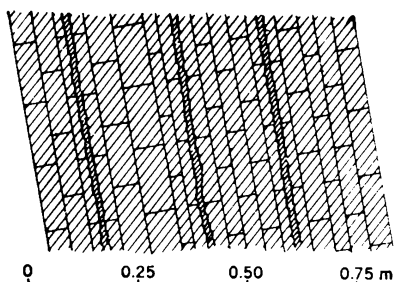


FIG. 205. Penetration of cleavage between layers of a section of Lower Cretaceous rocks on the Madnis-Ghele River in Soviet Georgia (after N. A. Rozanova).

At the same time, it is evident that the origin of principal flow cleavage is in no way connected with the sliding of beds over each other, as may be seen from an examination of the boundaries between the beds in a series containing cleavage. It will be noted that these surfaces, as a result of the cleavage, have a fringed or toothed look: the individual units of cleavage are so displaced relative to each other that the teeth of one bed engage the teeth of the other (Fig. 205). Such mutual engagement of the beds proves convincingly that there is no slippage between them in the formation of flow cleavage. From this, one can only conclude that the compression of the series of rocks during the formation of flow cleavage takes place without any bending of the layers and without any flow of the rock material within the beds. The entire series of folded beds is deformed as a unit. There is a slight discrepancy in the fact that the individual cleavage units are deformed to a somewhat different degree and, as a result, become slightly displaced relative to each other.

The general compression of a series of rocks leads to a change in the shapes of the geologic bodies composing it, including the beds. The latter decrease in dimension along the axis of compression and increase

in the direction of the axis of elongation. This determines the change in the thickness of beds, which is different in different places, depending on the position of the beds relative to the principal axes of deformation. If the axis of compression is horizontal and the strike of the beds is perpendicular to it, beds dipping at an angle of more than 45° decrease in thickness whereas those that dip at smaller angles increase in thickness.

The parallelism observed between the principal flow cleavage and the axial planes of the folds suggests that the overall compression is caused by the same forces that had earlier produced the folds by bending the beds and by a redistribution of the material in them.

It can be experimentally confirmed that flow cleavage is formed in the compression of a plastic body, regardless of the presence or absence of any bedded or folded structure in it. As early as the last century, Daubrée (1879) synthetically produced cleavage by compressing specimens of clays and metals. The latest experiments by I. V. Kirillova and Ye. I. Chertkova, in the Tectonophysics Laboratory of the Geophysical Institute, Academy of Science of the U.S.S.R., have shown that flow cleavage is easily obtainable by compressing homogeneous specimens of wax, paraffin, stearin, and ozocerite. Microscopic and X-ray analyses of the samples before and after deformation, revealed a reorientation of the crystals during this process, with their long axes perpendicular to the axis of compression. Thus, from the mechanical standpoint, synthetically induced cleavage is fully analogous to natural cleavage. Flow cleavage is observed above all in the more plastic layers. If the series of beds is not very intensively crumpled into folds, no cleavage at all may be observed in the more rigid beds. In more intensive dislocation, it "penetrates" into the rigid beds as well (such as sandstones and limestones) but in rougher form and more widely spaced. Only in very intensive crumpling of the layers does flow cleavage occur uniformly throughout all the rocks, regardless of their composition.

It must be noted that, although the mechanical environment in which principal flow cleavage originates is fairly clear, the mechanism of formation of the cleavage surfaces themselves is as yet unknown. The fact of a parallel orientation of the mineral particles is not alone sufficient to explain the formation in the rock, or in any plastic material, of closely spaced planes of weaker cohesion—which, in fact, the cleavage surfaces are. Why a rock splits into thin plates oriented normal to the direction of compression still remains unknown. The physical properties and the structure of the deformed material must play some part in this process. For instance, experiments with certain materials (petrolatum, gun grease) failed utterly to produce cleavage, whereas other materials (especially wax), under the same conditions, produced excellent cleavage structures. Which properties of the materials were the decisive ones

is still unknown. The problem of cleavage still awaits fundamental study, in which experimental methods will play a leading part.

The three elements of the mechanism of folding may be combined, in different cases, in various relationships and in varying succession. These variations as yet are little understood, and the conditions under which they appear have scarcely been determined. For this reason, only the briefest and most general remarks can be made here. When the folding is comparatively gentle, only shear cleavage is developed. These conditions involve primarily a bending of the beds and their sliding over each other. When bending is accompanied by a sufficiently intensive flow of the material within the bed, separate cleavage is developed in the individual beds. Principal flow cleavage develops when tighter folds have been formed and when further bending of the beds becomes difficult. A contributing factor is probably the high plasticity of the rocks, so that the resistance of the rock to general penetrative flow is no greater than the resistance of the beds to bending and slipping past each other. This stage of overall compression and flow, although it occurs after the folding, must be considered as its immediate continuation and development, in view of the close relationship between the direction of the cleavage and the position of the axial planes of the folds. The combination of this overall compression with the redistribution of the material within the bed probably leads to the formation of an inverted fan-shaped cleavage.

It has already been pointed out that the occurrence of bending is a manifestation of the mechanical independence of the bed and is due to the existence of boundary surfaces (bedding planes) between the beds. The redistribution of the material within the bed, though it demonstrates the interaction of the beds and their pressure against each other, nevertheless testifies to the mechanical independence that makes such flow within the bed possible. Finally, under overall compression and in the development of principal flow cleavage, the reaction of the entire dislocated mass to the mechanical forces is that of a single plastic unit wherein the individuality of the separate beds is almost completely lost.

These different stages of dislocation can occur in the same place, in the same formations, consecutively or jointly, during the same phase of folding. Some instances of the combination of stages have just been cited. The fact that various stages follow each other in time is indicated by numerous data: for example, rocks containing principal flow cleavage always have folds with beds thinned in the flanks and swollen in the crests and troughs. This proves that bending of beds and flow of the material within them took place before the general compression of the folded series. On the other hand, the S-shaped cleavage described earlier (in Chap. 8) suggests that a new stage of bending and slippage of the

beds over each other can take place after the overall compression. Indeed, the S-shaped cleavage can be most easily interpreted as principal flow cleavage sheared into segments corresponding to the individual beds and bent by the renewed sliding between them (see Fig. 57).

Such differences in the relative importance of the several elements of the folding mechanism in the process of deformation may be due to variations in the rate of deformation, which, in turn, affect the plastic properties of the material. It may also be affected by variations in the magnitude of the tectonic forces, by changes in the overburden as a result of erosion, and by other factors.

It must be said in conclusion that the questions discussed here have, generally speaking, not been sufficiently studied. One of the most urgent problems of geotectonics is the study of the entire complex of phenomena associated with the process of folding and the mechanical interpretation of this process.

MECHANICS OF THE FORMATION OF ORIENTED TEXTURES

It has been stated repeatedly in this book that oriented textures in rocks are the result of internal tectonic dislocations. Such textures originate primarily in folding; specifically, they determine flow cleavage. Oriented textures are also encountered in intrusive bodies and extrusive lava flows.

In all these cases, the basic mechanism is the same: differential movements of the particles within the rock (Billings, 1949; Pek, 1940; Fairbairn, 1949; Knopf, 1938; Sander, 1948–1950; Turner, 1948). This discussion will touch on certain problems connected with the mechanism of formation of oriented textures, on the basis of experiments conducted in the Tectonophysics Laboratory of the Geophysical Institute, Academy of Science of the U.S.S.R. (Belousov, 1950). It will be limited to the orientation of the *shapes* of the minerals—the orderly disposition of minerals that either possess or have acquired an elongated shape. As a rule, this orientation is manifested in the parallel positions of the long axes of many minerals throughout a more or less considerable volume of the rock. Optical orientation, meaning the orderly disposition of the optical axes of the crystals, regardless of their external shapes, is not considered, although some of the conclusions may throw light on that problem as well.

It has been seen that orientation of the mineral shapes can come about in two ways: either by a rotation of already elongated shapes until they become parallel, or by a deformation of the minerals and grains of the rock such that the originally random shapes become elongated in approximately the same direction.

First, let us consider the development of oriented texture resulting

from rotation of the grains. The originally randomly oriented elongated grains can acquire a parallel alignment by rotation only if there is a definite velocity gradient in the plastic flow occurring in the rock—if the velocity differs in different currents. In that case, an elongated grain, different parts of which are caught in streams of different velocities, will rotate in order to attain a position in alignment with the stream, with its long axis parallel to the current.

This has been experimentally demonstrated. A viscous fluid—a mixture of machine oil (2 parts) and gun grease (1 part)—was caused to flow down a gently inclined flume. Two kinds of flumes were used: one box-shaped, 12.5 cm wide, and one a semicylinder, 4 cm in diameter. Because of friction against the walls, the flow velocity on the sides was slower than in midstream. Minute wooden rods, 5 to 7 mm long and 1 to 1.5 mm thick, were dropped on the liquid at the upper end of the flume. Care was taken that the rods should fall either at random or with the majority of them across the current. Their number varied from 10 to 30. A streak of white talcum powder was sprinkled across the stream at the same spot. Several photographs were taken during the experiment. Direct observation, along with study of the photographs, revealed that in the course of the flow the streak of talcum across the current forms a bulge whose convexity points downstream and approaches a parabolic curve. The curvature increases gradually, from the edges to the middle, in such a way that the ends of the streak are curved first, whereas its middle part remains straight and perpendicular to the stream. The unaffected portion slowly grows shorter until the entire line becomes a parabola. Simultaneously, the rods are observed to rotate, first near the walls of the flume, then spreading toward the middle of the stream, but always more rapidly at the edges than in the middle, where the original orientation of the rods shows scarcely any change. It was noted that the completeness of their reorientation depends on the velocity of the flow: at a higher rate of flow, the rods become more perfectly oriented over the same distance of travel. Because the rotation rate was different in midstream and at the edges and because some of the rods that were already parallel to the stream continued to rotate, thus disturbing the orientation, statistical methods of evaluating the experimental data were used to achieve greater objectivity. Diagrams similar to a compass rose were constructed to indicate the rod orientation relative to the axis of the current at the beginning and end of the experiment. On the average, the terminal orientation of the rods was found to be more orderly than the initial. It may be concluded, therefore, that under analogous geologic conditions, such as those in which a magma carrying elongated crystals in suspension is intruded into a sedimentary series, orientation of the long crystals by rotation would develop first near the contact, gradually spreading

toward the middle of the intrusion. The depth to which this process of orientation would penetrate into the rock, and the completeness of the resulting orientation, would depend on the distance traveled by the flow, its velocity, and, obviously, its viscosity.

The mechanism by which oriented textures are formed through deformation of the grains themselves may be considered in the same manner. The same experimental apparatus of flume and flowing viscous fluid was used, except that, instead of the rods, white oil paint was sprinkled on the stream at the top of the flow, forming a series of fairly rounded spots on the oil's surface. The problem was to study the changes in the shapes of these drops during the course of the flow. The changes in the shapes of the drops during the experiment are illustrated by the series of photographs in Fig. 206. As they traveled downstream, the drops became more and more elongated but only at the middle of

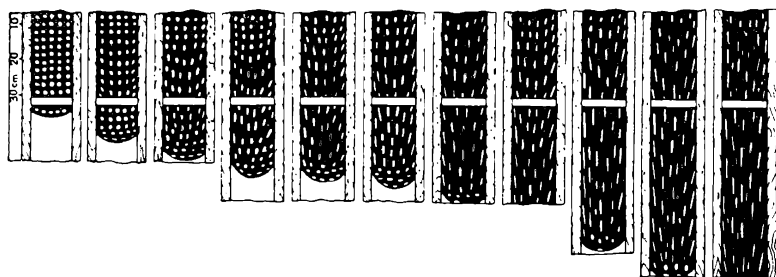


FIG. 206. Consecutive stages in the deformation of white paint drops on the surface of a viscous liquid, during the course of its free flow in a flume.

the current were the long axes of the rods parallel to the direction of the flow. Near the edges of the stream, the drops acquired an oblique orientation, with the projections of their long axes forming acute angles facing downstream. Finally, the elongation of the drops at the advancing center of the stream was parallel to its front.

What determines the changes in the shape of the drops? The flow in the flume is a free gravity flow. Under these conditions, a viscous liquid becomes elongated in the direction of the flow. At the edges of the stream, because of the velocity gradient, a shear deformation takes place, with a corresponding oblique direction of the long axis of deformation, reflected in the oblique orientation of the drops. In the advancing front of the stream, because of its growing convexity, the elongation of the drops is parallel to the front, thus determining their orientation.

The elongation must be connected with sliding along surfaces of shear oriented in accord with the maximum tangential stresses. Under an oblique light, which causes the surface of the liquid to appear glossy,

very fine and well-defined grooves may be discerned on it. These form two intersecting systems, as illustrated in Fig. 207. These grooves are shear surfaces corresponding to the direction of the maximum shear stress in various parts of the stream.

To determine the changes in drop orientation in a flow under pressure, the experiment was modified to cause the stream to flow in a horizontal flume by adding more liquid at one end. In attempting to spread out, the new liquid pushed the already present liquid along the flume, thus producing a pressure flow. Under these conditions, as illustrated by the series of photographs in Fig. 208, the deformed drops

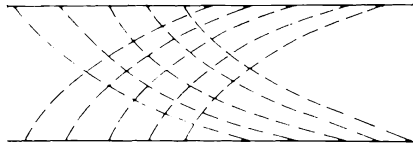


FIG. 207. Distribution of shear lines on the surface of a viscous fluid flowing down a flume. Current flows from left to right.

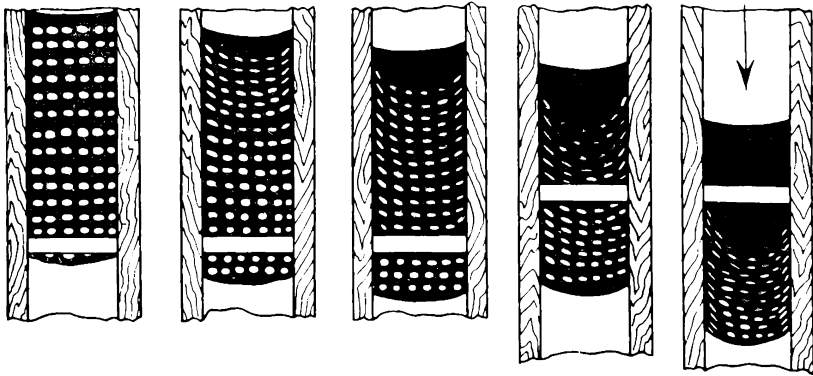


FIG. 208. Successive stages in the deformation of oil paint drops on the surface of viscous liquid flowing in a flume, under pressure from the direction of the arrow.

formed arcs that were convex downstream, with the greatest deformation of the individual drops along the edges, near the source of the pressure. Thus in free flow in an empty space (which occurs in nature, perhaps, only in lava and mud flows), the orientation produced by deformation of the grains must be reflected in the disposition of the long axes of the grains parallel to the direction of flow in the middle of the stream and at an angle to it near the edges. In a pressure flow, which has more analogies in nature, the deformed minerals develop an arched orientation.

Similar experiments could be carried out with various plastic materials of high viscosity (heated paraffin, gun grease, petrolatum). To

observe the changes in the internal texture, spherules of the same or different material may be placed in the substance. Spherules of the same material are made visible by sprinkling colored powder over them. The deformation of the spherules, as seen in cut sections, reflects the deformation in the enclosing medium and indicates changes in its texture.

If a cylindrical specimen of heated paraffin is compressed in the direction of its axis, allowing the material to flow freely in all directions normal to the pressure, ozocerite spherules embedded in the paraffin are flattened and turned into ellipsoids. Under these conditions, the ellipsoids appear to be biaxial: only in the sections parallel to the direction of the pressure does one see, instead of disks, ellipsoids, with their long axes normal to the pressure. In sections perpendicular to the direction of the pressure, the flattened spherules preserve their circular shape. This experiment has thus produced a model of a flat internal

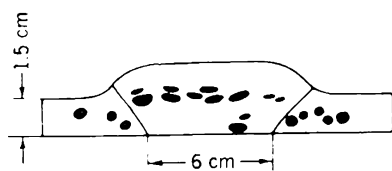


FIG. 209. Deformation of spherules in the squeezing out of a cylinder under a sand overburden. Mixture of petrolatum and paraffin.

structure: one in which the flat minerals lie parallel to each other, although their long axes are haphazardly disposed in the plane of orientation. The origin of a flat texture in rocks is evidently determined by the ability of the compressed rock to spread in all directions perpendicular to the pressure. If, in the process of compression, the texture is to become not only flat but linear, the flow and extension of the rock should be limited to a single direction.

A number of examples may be cited of artificial orientation in textures, achieved by deforming ozocerite spherules contained in a plastic medium under the conditions of certain particular kinds of deformation. Figure 209 shows the results of an experiment in which the central part of a horizontally stratified plastic medium (a mixture of petrolatum and paraffin) was squeezed upward. This was done by pumping in more of the same material from below, through a cylindrical vent. The sample itself was contained under an overburden of sand; because of the resistance of this overburden, the squeezed-out portion was flattened on top and contained on all sides by shear surfaces. Its horizontal spread was reflected in a corresponding deformation of the spherules of the same material, which became ellipsoids with their long axes horizontal.

In the formation of a dome in very soft plastic layers by upward pressure from below, the spherules became elongated parallel to the curvature of the layers (Fig. 210). When the bending of the layers is caused by tangential pressure, the orientation is different: the spherules become ellipsoids elongated radially in relation to the fold. When the

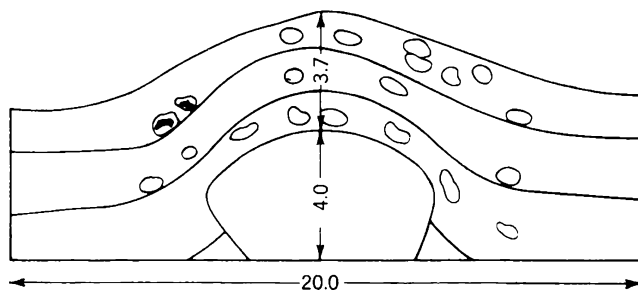


FIG. 210. Deformation of spherules in the formation of a dome.

fold is tightly compressed, the long axes of the ellipsoids are parallel to its axial plane.

The mechanism of the formation of oriented textures under various conditions of movement of the plastic medium can be studied by this method; conversely, when one is confronted with a known texture in a rock, the conditions of its formation may be reconstructed. Oriented textures have also been produced directly in rock samples, such as marble, subjected to high pressures.

CHAPTER 21

FORMATION OF CONTINUOUS FOLDS IN TIME AND SPACE. FOLDED ZONES AND PLATFORMS

SLOW CONTINUOUS FOLDING VERSUS BRIEF PHASES

Some investigators believe that all folding, including continuous folds, results from slow and gradual movements of the earth's crust. Folding, in their view, is conceived as a continual process varying from epoch to epoch in intensity and in the places in which it is manifested. This is the so-called theory of constant folding. Its adherents are N. S. Shatskiy, (1937*b*), D. V. Nalivkin (1936), V. Ye. Khain (1938) and others, and Bärtling, Packelman, Novak, and certain other workers outside the Soviet Union.

Another (and larger) group of investigators maintains that the process of continuous folding occurs in spurts or phases separated by periods of quiescence. Each phase must be considered as a rapid (in the scale of geologic time) folding of the beds. Its duration is probably on the order of several hundred thousand years—so short, compared with the time intervals recorded in geologic sections, as to be marked in the earth's crust as an almost instantaneous event. The age of a phase is determined, it is well known, by angular unconformities.

The adherents of the theory of constant folding reasoned as follows. First, they cited the instances of a thinning of the beds in the crests of anticlines and of their thickening in the troughs of synclines; this was supposed to be evidence of folding simultaneous with the accumulation of sediments. Second, they held that the presence of angular unconformities does not necessarily prove the episodic nature of the folding, since such a break would mark only a vertical uplift of the earth's crust rather than the time of the folding; nor does it signify anything about the brevity of the process. The difference in occurrence of the beds above and below the unconformity is explained as a measure of the magnitude of the folding movements that took place during the break.

This might have been so if in continuous folds a thinning of the beds

had been observed in the anticlines, as is the case in discontinuous folds. But careful study of the instances of such variation in the thickness as are found in the literature leads to the conclusion that the examples cited are obvious misunderstandings. A majority of them turn out to be typical discontinuous, rather than continuous, folds. This is true of the structures in Albania and Dalmatia that have been mentioned by foreign authors, or of Kabristan (southeastern Caucasus) and the Kura depression as described by Soviet investigators (Shatskiy, 1937*b*). Their references to a decrease in thickness in the crests of the anticlines, accompanied by a change toward coarser facies, are quite correct, but such instances are not unusual, in that the folded structures in those areas are definitely discontinuous.

In other instances of continuous or intermediate folding, the references to such changes in the facies and thickness have proved to be erroneous. Such is the case in the folded Ruhr basin, where the signs of facies and thickness changes were first interpreted as evidence of concurrent sedimentation and folding, and of the long duration of the latter (Bärtling, 1927); subsequently, this interpretation was rejected (Keller, 1940). The same was true of the Donets basin, where N. S. Shatskiy (1937*c*) pointed out a thinning of the beds in the anticlinal crests, only to be refuted by A. S. Shirokov (1938).

Thus there is now not a single verifiable instance of continuous folding with the alleged facies and thickness changes suggesting a slow folding process. In hundreds of sections through various folded zones, no connection has been observed between the distribution of the primary thicknesses and the facies, on the one hand, and the individual folds, on the other. Secondary mechanical factors, as has been pointed out, tend to increase the thickness in the crests and troughs of the folds.

To this the proponents of the constant folding theory object that the succession of beds containing these facies and thickness changes has been removed by erosion and that such changes therefore cannot be seen. But it is highly unlikely that erosion would have removed the upper strata everywhere, without exception.

The decisive instances are those in which an angular unconformity is associated with a very short stratigraphic break and with a small amount of erosion. In such cases, the absence of facies and thickness changes in the folds can only mean folding of short duration, no longer than that of the stratigraphic break. Such cases are very common. V. Ye. Khain (1950*c*) mentions a 35° unconformity between the Upper Tithonian and the Lower Valanginian stages in the Gil'ginchay River basin in the southeastern Caucasus; in the Lesser Caucasus, near Kirovabad, a sharp angular unconformity has been noted between the Campanian and Maestrichtian; in Abkhaziya, near the village of Pskha, the Lower Cretaceous unconformably overlies an overturned fold of the Bathonian beds;

in the Crimea, a considerable unconformity is known to exist between the Lower and Upper Callovian, in the Pamirs, there is an unconformity between the Callovian and Oxfordian; and there is another between the Lower and Middle Tournaisian in the western Tien Shan. In the non-Soviet literature we learn that in Lower Silesia the top of the Lower Carboniferous forms an angular unconformity with the basal beds of the Middle Carboniferous; in the Saar there is an angular unconformity between the youngest beds of the Lower Old Red Sandstone and the oldest beds of the Upper Old Red Sandstone. In the Pyrenees, the basal Cenomanian beds unconformably overlie the Upper Albian rocks. In the Cisalpine zone, the Lower Pliocene (Pontian) is folded, whereas the Middle Pliocene (Piazzentian) is undisturbed. In no case do the field data clearly indicate a very long duration of the formation of continuous folds. Many data contradict this idea and indicate rather that the folding is of short duration and that it is closely related to uplift of the earth's crust (in cases of an angular unconformity with a short stratigraphic break).

This last circumstance is extremely important. The constant folding theory assumes that there is no regular association between the oscillatory movements of the earth's crust and folding and that these two types of tectonic movements occur independently of each other. The present writer's contention is quite the opposite: he attempts to see unity in all the forms of tectogenesis, and the totality of all existing data suggests that this unity indeed exists.

The theory of constant folding creates methodologic difficulties in preventing the use of angular unconformities to decipher the tectonic history of a region and to establish the epochs of critical significance in the formation of its structure. Thus the historical roadmarks are lost, and the tectonic process spreads out indefinitely, becoming a slow evolution of structural forms upon which some extraneous oscillations of the earth's crust are superimposed.

V. Ye. Khain (1950c), in contending that tectonic processes occur both constantly and intermittently, now believes that the intermittent nature of the growth process—the individual spurts or phases—is more clearly manifested in continuous folding than in discontinuous. He wrote:

In discontinuous folding, the intensity of the growth of the folds shows less variation in time. Therefore the entire process is to be understood as a slow and gradual development. On the other hand, in the formation of linear folds, the *decisive* [italics mine: V. V. B.] role is played by intensive movements simultaneous with the general uplift, which *camouflage* [italics mine: V. V. B.] the effect of the slow growth of the folds in the interval between the uplifts.

At the same time, Khain underscored the connection between a sharp intensification of continuous fold formation (phases of folding) and the

oscillatory movements specifically, the short-lived uplifts of the earth's crust.

N. S. Shatskiy (1951), in his paper directed against "neocatastrophism" and "Stilleanism" (his names for the views held by the adherents of the idea of brief phases in continuous fold formation), admitted the existence of phases of folding as periods of "disproportionate folding leading to abrupt changes in the general structure of a given segment of the earth's crust." The following sentence explains what he means by "changes in the general structure": "In the course of a phase the folds become more complex, their flanks become steeper and steeper, and secondary deformations develop, complicated by thrusts, cleavage, etc." When one adds that Shatskiy admitted a postphase interruption in the formation of folds, the quoted characteristics do not differ in quality from the generally accepted definition of a phase.

Unfortunately, there are no data on the absolute duration of the phases of continuous fold formation. On the basis of the brevity of a number of known stratigraphic breaks associated with angular unconformities (cited above), it is known only that these phases were short-lived; but whether they lasted for hundreds of thousands or for several millions of years, it is still impossible to determine.

Thus one arrives at the conclusion that the two basic morphologic types of folding also differ in their origins in time: if discontinuous folds are formed during the course of a slow and long-lasting vertical movement of the earth's crust, accompanied by simultaneous deposition and punctuated by separate, more intensive pulses, continuous folds are a result of relatively brief phases of folding movements, separated by much longer periods of quiescence.

EPOCHS AND PHASES OF FOLDING

A study of the angular unconformities in various folded zones reveals that the phases of folding are not haphazardly scattered in time: rather, there are certain epochs in the earth's history with which these phases are associated. Thus one may speak of epochs of folding.

It is known that the formation of continuous folds is localized in geosynclines. Accordingly, any discussion of the epochs of continuous fold formation must take account of their relationship to geosynclinal zones. But within the latter (meaning, of course, only "active" geosynclines), these epochs are manifested almost everywhere; their effects, therefore, encompass the entire area of the continent.

A large number of these epochs of folding took place in the Precambrian. From the end of the Precambrian to the present time, three epochs of folding are recognized. The first occurred in the Silurian period, the second in the Middle and Late Carboniferous and in the

Permian, and the third mainly in the Tertiary and partly in the Cretaceous. These three epochs of continuous fold formation correspond to the three geotectonic cycles previously distinguished on the basis of a study of the oscillatory movements of the earth's crust: the Caledonian, the Hercynian, and the Alpine periods of folding. It must not be thought that no phases of folding occurred outside these epochs. Such phases are known to have taken place, but their manifestations are less frequent and less intensive than those of the phases associated with epochs of folding. Since the epoch of folding, as one can see clearly, occurs toward the end of the geotectonic cycle, these other phases evidently occur in the first half of the cycle. Moreover, even within a single cycle, the phases are of different magnitudes, exerting various degrees of influence on the formation of the folded structure of the area. The succession of stronger and weaker phases is different in different places, although it is observed that the phases become weaker near the end of an epoch.

In this connection, in view of the weak phases of folding at the beginning of the cycle, the stronger folding at the start of the epoch of folding, and again the weaker phase at its end, it is possible to divide each geotectonic cycle into *preliminary*, *main*, and *terminal* phases of folding. It will be seen that this division is determined not only by the relative magnitudes of the phases but also by their location within the geosyncline, their type, and their relative significance in the formation of the folded zone as a whole.

Generally speaking, although the epochs of folding are universally distributed, the same cannot be said of their particular phases. Stille (1924) attempted to prove the universality of the "orogenic" phases by setting up a "canon," or list of phases, which he believed to be applicable to any geosynclinal region of the earth. Actually, however, it was soon shown that this rigid distribution, in the stratigraphic time scale, of a definite number of phases purporting to be universal is very far removed from reality. In various parts of the U.S.S.R. phases of folding have been found that were not foreseen by Stille's canon. Thus attempts were made to supplement the canon with new phases. The Labinsk, Chegem, Trialet, Kabristan, eastern Caucasus, and Baku phases were added in the Caucasus; the Donets phase in the Donets basin; the Chatkal' phase in the Tien Shan; the Pamir and Issyk phases in the Pamirs; the Salair phase in western Siberia, etc. Within ten years Stille himself was forced to augment his canon by ten new phases.

It soon became clear that it did not matter whether the canon consisted of twenty or thirty phases; the very attempt to construct a dogmatic register of universal phases was basically incorrect. It has since become clear that even within the limits of one geosyncline—even a single intrageosyncline—the phases of folding are very unevenly dis-

tributed in area. Each phase occurs locally and is not observed elsewhere. In different parts of the same intrageosyncline, the simultaneous manifestations of folding differ greatly in their intensity. It will be seen that an orderly migration of the phases of folding takes place in the territory of a geosyncline.

If one compares geosynclines that are distant from each other, such a great divergence in the time of occurrence of the several phases of folding appears that there can be no question of a single universal list of phases. Even where, in a cursory investigation, the phases of folding appear to be contemporaneous, a more detailed study as a rule reveals considerable divergences in time. Many of Stille's phases have thus lost their definite position in the stratigraphic time scale and have come to embrace a number of separate phases occurring at different times and in different places. Other phases, under closer scrutiny, seem to break up into a series of phases following each other within a short, but still geologically perceptible, time interval. In the Rocky Mountains, for example, the onset of the so-called Laramide phase, originally placed at the boundary between the Cretaceous and the Eocene, turned out to encompass, instead of a single phase, several phases extending in time over the end of the Cretaceous and the beginning of the Paleogene; some of these phases are observed in certain areas and others elsewhere, and none is encountered over more than a very limited territory. If one seeks a Laramide phase in the Caucasus, one will find it at the beginning of the Danian stage in the northwestern Caucasus, whereas in Dagestan angular unconformities have been found at the middle of the Danian stage as well as between the Lower and Middle Eocene. In the southeastern Caucasus the unconformity occurs between the Lower and Middle Paleocene, whereas in the southern part of the Talyshskiy range the Middle Eocene lies unconformably upon the Paleocene (Khain, 1950c).

Thus the danger of an uncritical use of Stille's canon becomes evident. On many occasions, geologists, feeling themselves in duty bound to find and recognize a particular phase of folding merely because it appeared on Stille's register, sought refuge in artificial devices. Mere stratigraphic gaps brought about by oscillatory movements were included among the phases of folding; even interbeds of conglomerate were taken as "manifestations of orogenic phases." Even more dangerous are attempts to use the canon for stratigraphic purposes, when a geologist "determines" the age of a rock series on the basis of the supposedly necessary stratigraphic position of the angular unconformity and the break.

The only correct approach is a painstaking and unprejudiced study of what the geologist sees before him. In any region, the real and not the imagined history of its continuous fold formation must be recon-

structed from the angular unconformities in the section through its sedimentary rocks. Only over a very limited area, and only as a preliminary, may these unconformities be used as a stratigraphic criterion. It must be kept in mind that, even over a limited area, the distribution of the phases of folding may differ in time.

In the Caledonian cycle, the intensive phases are concentrated mainly close to the boundaries between the Ordovician and the Gotlandian, between the Ludlovian and the Downtonian, and between the Gotlandian and the Devonian. In the Hercynian cycle they occur between the Upper and Lower Devonian, at the base of the Lower Carboniferous, and between the Carboniferous and the Permian. In the Alpine cycle, the phases appear before the Upper Jurassic, near the beginning of the Tithonian stage, at the boundary between the Lower and Upper Cretaceous, at the end of the Cretaceous and the beginning of the Paleogene, between the Eocene and Oligocene and the Oligocene and the Miocene, and between the Miocene and Pliocene, at the end of the Pliocene, and the beginning of the Quaternary. Either a single phase or several consecutive phases in different places may appear in any of these instances.

By means of his canon, Stille introduced his own nomenclature for the several phases into general usage. His names are chiefly geographic: Ardennian, Bretonian, Saalian, Austrian, Pyrenean, etc. Upon discovering new phases, geologists usually gave them geographic names (Adygey, eastern Caucasus, Labinsk, Pamir, etc.). To avoid misunderstandings—which are quite possible, as previously explained—a phase nomenclature based on the stratigraphic position of the phases is recommended. Moreover, if it is definitely established that a phase occurred immediately before a stratigraphic subdivision, it should be named after this subdivision, using the prefix “pre-” (pre-Baku, pre-Meotisian, pre-Downtonian, etc.). If, on the other hand, the break is of long duration and the phase could have taken place during a long time interval, the prefix “ante-” might be used. If, for instance, a break associated with an angular unconformity encompasses the time from the Early Jurassic to the Tithonian, the ante-Tithonian phase may have taken place any time between the Early Jurassic and the Tithonian. In certain instances, finally, it will be expedient to add the prefix “post-.” This will be the case when the suite unconformably overlying a folded stratum is missing and the upper limit of the age of the folding is thus indefinite. When, for instance, the youngest folded beds belong to the Carboniferous, with no identifiable deposits above them, a post-Carboniferous folding may be referred to.

In cases in which a geographic name is the most convenient one, the age of the phase in question should be defined as precisely as possible.

FOLDED ZONES AND PLATFORMS

Beside the association of fold formation with definite epochs, it has been observed that there is a regional concentration of the folding dating from a particular time. The folding of each epoch is closely concentrated in certain regions and zones, thus giving rise to zones of Caledonian folding (Caledonian structures), zones of Hercynian or Variscian folding (Hercynian or Variscian structures), and zones of Alpine folding (Alpine structures). An example of Caledonian structure is the Scandinavian mountains, in which the last folding took place in the Silurian period; an example of Hercynian structure is the Urals, with their Late Paleozoic folding; and, finally, an example of Alpine structure is the Caucasus, whose main phases of folding are associated with the Tertiary period.

More ancient Precambrian folded zones have also been distinguished. Some writers describe a separate zone of Cimmerian or Mesozoic folding (Cimmerian or Mesozoic structures). The question of whether there was an independent Mesozoic geotectonic cycle has already been discussed above; in these cases, the present writer would prefer to speak of an Alpine cycle, whose stages all begin somewhat earlier, since the most intensive phases of folding occurred not in the Cretaceous or Tertiary but in the Jurassic.

The folded zones of different cycles overlap each other, so that in a single locality the rocks may have been subjected to repeated intensive folding. The folded zones coincide with the geosynclines of the corresponding age and are the result of the latter's structural development.

Outside the folded zones are areas that during the given cycle did not undergo formation of continuous folds; these are the same as the platforms. A definition of platforms was given earlier, on the basis of the regime of oscillatory movements of the earth's crust. A platform may also be defined in other terms, however, as a definite structural complex characterized by the development of antecises and synecises and by the absence of continuous folds. This definition was mentioned in Chap. 9. It is obvious that these two definitions embody different senses of the word: historical (referring to the regime of movements) in the first case, and structural in the second. In the first definition, platforms are opposed to *geosynclines*, with their intensive and highly contrasting oscillatory movements and active manifestation of other tectonic processes; in the second definition, they are opposed to *folded zones*. The folded zone is characterized fundamentally by its extensive development of continuous folds.

We shall attempt to clarify the origin of platforms in terms of their structural definition. Caledonian platforms are areas that were not involved in the formation of continuous folds during the Caledonian cycle. Similarly, the areas that were not involved in the formation of

continuous folds in the Hercynian cycle are Hercynian platforms, and, finally, the Alpine platforms are the areas that were not drawn into the continuous folding of the Alpine cycle. All these definitions hinge on the formation of continuous folds. It has already been seen that discontinuous folds not only may be formed on platforms but are, in fact, very widespread on them.

During the Caledonian cycle, the Russian plain was a Caledonian platform; that is, Caledonian continuous folds were not formed on it. In the Hercynian cycle this platform expanded at the expense of the adjoining Scandinavian mountains: these were the site of Caledonian folding movements, but they did not undergo Hercynian continuous folding and were, consequently, part of the Hercynian platform. In the Alpine cycle, finally, this same platform also expanded eastward at the expense of the Urals and western Siberia, which had been a Hercynian folded zone but where there was no formation of continuous folds in the Alpine cycle.

Thus the Alpine platform encompasses all the territory of Western Europe outside the limits of the Alpine folded zone, as well as the Russian plain, the Urals, and also western and central Siberia as far east as the Verkhoyansk Range. In the Hercynian cycle there were two independent platforms—the Russian and the Siberian—separated by the folded zone of the Urals; in the Alpine cycle, however, they were united into a single great European-Siberian Alpine platform.

In answer to the question of what the Russian plain has been, in regard to its structure, since the beginning of the Paleozoic, it may be said that this area formed part of the Caledonian and the Hercynian and the Alpine platforms but that the areas and the outlines of these platforms were different: the Alpine platform was considerably greater than the Hercynian, and the Hercynian greater than the Caledonian. This increase in the size of the platform during the course of the cycles is a general rule. The platforms expand at the expense of the geosynclines from cycle to cycle, and the platforms of succeeding cycles must therefore be greater than those of the preceding ones.

Moreover, each platform is a mosaic of parts of older folded zones. Taking the entire vast European-Siberian platform of Alpine age as an example, the southern boundary of which is the same as the boundary of the Alpine folded zone (that is, the boundary of the Alps, the Carpathians, the Balkans, the Caucasus, the Kopet Dag, the Pamir, eastern Transbaikalia, and the Verkhoyansk Range), one can clearly discern this mosaic. In fact, Sweden, Finland, and the Russian plain, on the one hand, and central Siberia, on the other, are both areas that were crumpled into folds as early as the Precambrian. These are the remnants of Precambrian folded zones that were platforms during the Caledonian cycle.

Norway, England, France, Spain, and almost all of Germany and the

northern Tien Shan and the Baikal area are parts of Caledonian folded zones. Finally, central Spain, almost all of France, southern and central Germany, the Urals, western Siberia, and the central Tien Shan are parts of Hercynian folded zones that were included in the Alpine platform. Consequently, one may distinguish areas of the Alpine platform that were formed on folded basements of various ages and speak of an Alpine platform on a Precambrian folded basement, or on a Caledonian or Hercynian basement.

Because of the fact that platforms are situated on earlier folded zones, their internal structure always contains two complexes of rocks: a lower complex crumpled into continuous folds and formed under geosynclinal conditions, and an upper complex that is either completely undisturbed or dislocated only by discontinuous folds formed under platform conditions that replaced the earlier geosyncline. The lower complex is usually called the *basement*, or *folded basement*, of the platform. The age of its folds may vary, as has been seen, but since in the Precambrian geosynclines existed everywhere and continuous folds were consequently formed everywhere, all platforms, whatever their age, will in every case have a Precambrian folded basement.

It must be admitted that, although the classification of folded zones by age is generally accepted, this is not so with the distinction between platforms according to the age of their structures. On the other hand, it is impossible to avoid classifying platforms by age, since the shapes and areas of platform regions vary considerably from cycle to cycle. Tectonic movement, moreover, is constant and universal, and the structure of an area that is under platform conditions changes over the course of several cycles. Thus, for instance, the Russian platform did not have the same structure in the Alpine cycle as it had in the Hercynian, and in the latter cycle its structure was again different from what it had been still earlier. For this reason it is not sufficient merely to refer to the Russian platform: to define the structural concept, one must indicate which cycle of the platform one is discussing.

The age of a platform is sometimes designated by the age of its folded basement. In this case the territory of the Russian plain will be called a Precambrian platform, since the folded basement here is of Precambrian age. In the same sense, the Scandinavian mountains must be considered a Caledonian platform, western Siberia and the Urals a Hercynian platform, and so forth. This principle of defining the age of a platform, however, is erroneous. In the first place, it clearly involves a contradiction in terms: a Precambrian platform should, naturally, be one that existed in Precambrian times and not one formed after the Precambrian. In the second place, misunderstanding inevitably arises in the juxtaposition of the corresponding folded zones and platforms: the Scandinavian mountains, according to the terminology criticized

here, are at once a Caledonian folded region and a Caledonian platform; the Urals likewise belong to both the Hercynian folded zone and the Hercynian platform, etc. Finally, in this method of designating platforms by age, with the transition in an area from geosynclinal to platform conditions, every further structural development there is suspended. For this reason, it is not the time of formation of the given platform structure that is significant but the time at which the tectonic movements supposedly came to an end.

It cannot be denied, however, that in certain cases it is important to indicate the time when the platform was formed; that is, when the geosynclinal conditions gave way to platform conditions in the given territory. Here it is convenient to use the prefix "epi-" with the name of its last geosynclinal cycle and, for example, to call a platform epi-Hercynian if it was formed after the end of the Hercynian geosynclinal cycle in that region—that is, since the beginning of the Alpine cycle (beginning of the Mesozoic). Epi-Caledonian platforms were formed after the end of the Caledonian cycle (since the beginning of the Hercynian cycle), and epi-Proterozoic platforms after the end of the Proterozoic cycle, during which they were still under a geosynclinal regime. Thus an epi-Caledonian Alpine platform will be a structure formed during the Alpine cycle in a platform existing since the beginning of the Hercynian cycle.

CHAPTER 22

THE DEVELOPMENT OF FOLDING IN GEOSYNCLINES: ITS RELATIONSHIP TO OSCILLATORY MOVEMENTS

The association between the regime of oscillatory movements and the formation of folds is evident. In its most general form, this relationship is reflected in the difference between folding on platforms and in geosynclines. Discontinuous folds are developed on platforms. In geosynclines the folding processes are various: intensive continuous folds are widespread, but, in addition, there is also the formation of intermediate and discontinuous folds.

The development of discontinuous folding on platforms was examined earlier; this chapter will discuss the folding in geosynclines. The conditions determining the formation of continuous folds will be examined along with those that govern folding of other types in geosynclines, since there is, as we have seen, a unity common to all the processes of folding.

MIGRATION OF THE PHASES OF FOLDING IN GEOSYNCLINES

The development of folding within a geosyncline occurs in close association with the development of oscillatory movements. The most generally noticeable relationship appears in the fact that the boundaries of the folded zones of a given age correspond to the boundaries of the geosynclines of the same age (the same cycle).

It was said earlier that geosynclines frequently take peculiar shapes. For instance, they may take the form of ovals; in such cases the folded zones that repeat the outlines of the geosyncline form similar ovals, either isolated or adjoining. This is the picture presented by the Alpine folded zone in Europe.

This touches upon a question that is frequently misinterpreted. Beginning with Suess (1875), many investigators have said that the ir-

regular curved outlines of a folded zone are determined by the outlines of the rigid blocks whose movement toward each other has produced it. This interpretation completely ignores the preceding subsidence of the geosyncline and the fact that the folded zone merely follows the outlines of the geosyncline. To explain the contours of a folded zone, the causes of the observed shape of the geosyncline must be understood.

Another relationship between folding and oscillatory movements is manifested in the fact that for each stage in the development of oscillatory movements, there is a corresponding stage in the formation of the folded structure.

A study of folded zones reveals not only different degrees of intensity of folding in different places and different inclinations of the folds but also differences in the times of occurrence of the unconformities. In different areas the folding appears to have taken place during different phases although in the same cycle (considering only the rocks deposited during a single cycle). In some parts of the zone the folding has taken place earlier and in other places later. This lack of synchronization in the formation of folds of the same cycle, and in the same geosyncline, is extremely interesting, especially in view of the fact that the migration of the phases in time and space is subject to certain fully determinable laws.

Earlier in this book, the phases of folding within the cycle were divided into preliminary, main, and terminal. Study of the corresponding angular unconformities reveals that these phases are developed over different parts of the geosyncline and that each group of phases belongs to a definite stage of the oscillatory movements: the preliminary phases are developed before the inversion of the geotectonic regime and appear in the intrageanticlines; the main phases are most intimately associated with the inversion and are developed in the central uplift; with the expansion of the central uplift, the area of folding expands with it, away from the axis of the uplift, until the terminal phases are developed in the foredeeps and intermontane basins. In addition, the shapes (types) of the folds also change. The preliminary phases associated with the preinversion intrageanticlines produce intermediate (box-shaped) or even discontinuous folding. Generally speaking, it has turned out that, although the formation of discontinuous folds is a long-lasting and gradual process, it varies greatly in time, so that the concept of phases may be applied to this type of folding as well. This concept is particularly useful as applied to discontinuous folds formed in geosynclines.

The majority of factual data relate to the main and terminal phases. The origin of intensive folding primarily in the interior parts of the geosynclines (in the central uplifts) and its later displacement toward

the periphery of the geosynclinal zone have been known for a long time in the literature as the *migration of phases*.

Data bearing on the preliminary phases are scarcer. This is due to the fact that, in the second half of the cycle, the intrageanticlines become basins filled with younger sediments, concealing the parts of the section that contain the more ancient phases.

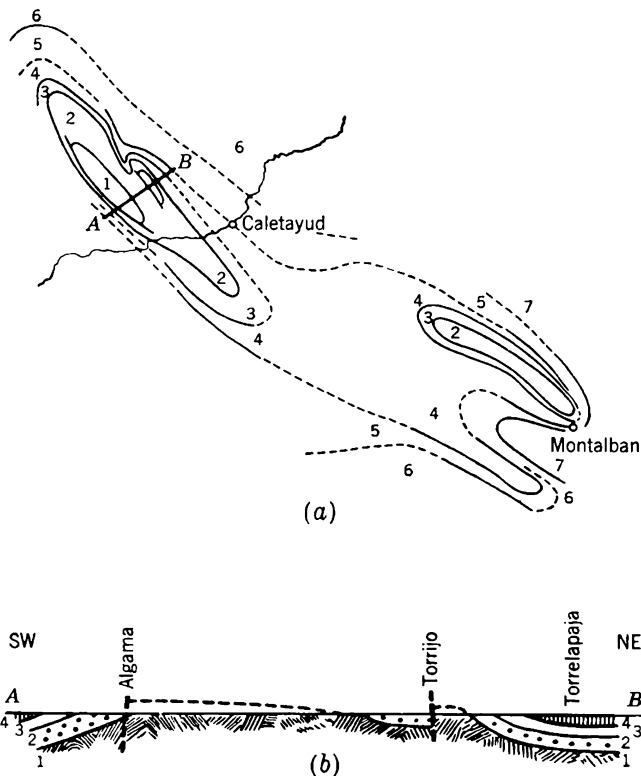


FIG. 211. Pre-Late Cretaceous paleogeologic map of the Montalban ridge in the Celtiberian zone (a) and its cross section (b): (1) Paleozoic; (2) Lower and Middle Triassic; (3) Upper Triassic; (4) Rhaetian; (5) Jurassic; (6) Hauterivian; (7) Urgonian.

Examples are found in the Celtiberian geosyncline, the Greater Caucasus, and the western Alps. In the Celtiberian geosyncline (Richter and Teichmueller, 1933), there was an uplift before the Late Cretaceous in the Montalban intrageanticline, bringing the Upper Cretaceous into an erosional contact with the underlying older beds. A paleogeologic map of the eroded surface shows the Upper Cretaceous resting on various horizons, 'in such a manner as to suggest a previous dislocation resulting in domical folds locally broken by high-angle thrusts (Fig. 211). Thus there was a phase of weak folding before the Late Creta-

ceous. The resulting folds are close to the discontinuous type. These comprise all the preinversion phases of folding known in the Celtiberian zone.

In this region an inversion took place before the Paleogene; with it was associated the main, or pre-Paleogene, stage of folding, in which all the beds in the central uplift, up to and including the Upper Cretaceous, were folded. This phase did not appear in the foredeeps, where the sediments, including the Eocene and Oligocene, were folded during the pre-Oligocene and pre-Pliocene phases, which are the terminal phases. It is hard to say whether these phases were confined to the foredeeps or also took place in the central uplift. If the latter is true, the beds of the central uplift were involved in a larger number of phases than those of the foredeeps, and there was a postinversion expansion of the area of folding from the centers of the intrageosynclines toward their periphery. If, on the other hand, the terminal phases were confined to the foredeeps, there was a displacement of the site of folding in the same direction. In any event, the younger formations in the depressions are folded less tightly than the more ancient folds of the central uplift, and the folding is of the intermediate (comb-shaped) and discontinuous types.

In the Greater Caucasus, the development of folding against the background of the oscillatory movements has been studied in detail. Prior to the inversion, which is known to have occurred here during the Oligocene, the following folding phases took place during the Alpine cycle: the pre-Late Liassic, pre-Callovian, pre-Tithonian, pre-Valanginian, pre-Late Cretaceous, and pre-Maykopian. On the basis of an analysis of the distribution of the areas of these phases, it can be said that in all the above instances the folding movements were confined to the intrageanticlines and were fully absent in the intrageosynclines. The data on this question are presented in another book by the present writer (1938-1940).

In the Caucasus, the preliminary phases were of low intensity; they were developed over certain limited areas and had only very little effect on the final folded structures of the Caucasus. These belong to the category of intermediate (boxlike) and discontinuous folding.

An intensive formation of continuous folds begins at the moment of inversion. The pre-Chokrakian, pre-Middle Sarmatian, and pre-Meotisian are the main phases in the Caucasus. These first appeared in the newly elevated central uplift and then spread over the entire main Caucasus range and were of great intensity.

They were followed by the pre-Pontian, pre-Produktivnaya, and pre-Akchagyl' phases, several Apsheron phases, and the pre-Baku and post-Baku phases of folding. These terminal phases, especially the later ones, had a considerably wider distribution. The folding involved not only

the central uplift—where the terminal phases also appeared—but the adjacent depressions as well: the foredeep in the north and the intermontane basin in the south. At that distance, however, the intensity of the folding was dampened, and its manifestation was primarily of the intermediate or discontinuous type. As in the previous case, the central uplift was folded more intensively than the adjacent depressions, as a result of the greater number of phases occurring in it and of the decrease in the intensity of the folding as it advanced from the central uplift to the depressions.

All the phases of folding were accompanied by uplifts of the earth's crust. The preliminary phases are associated with brief and sharp uplifts in the intrageanticlinal regions, leading to erosion of some of the deposits and to unconformities between these and the later beds. The postinversion phases took place against the background of the growth of the central uplift, although each phase was accompanied by a supplementary acceleration of the elevation of the earth's crust, as if it had been thrust upward.

In the southern part of the western Alps, the inversion took place before the Eocene. A single preliminary phase of folding in that region is well authenticated: this is pre-Late Cretaceous. This appears only in the Briançon geanticline and is associated with a time of increased intensity in its uplift. If, both before this phase and after it, the Briançon zone differed from the neighboring intrageosynclines in the smaller degree of its subsidence and the thickness of its deposits, before the Late Cretaceous an absolute uplift of considerable magnitude must have taken place there. As a result, the older rocks underwent considerable erosion, so that the transgressive Upper Cretaceous does not begin with its lower strata. The folding phase is marked by the angular unconformity at the contact between the Upper Cretaceous and the underlying rocks.

The main phases are pre-Eocene and pre-Oligocene. The first is developed exclusively over the main intrageosyncline—the Pennine—and coincides in time with the inversion and the formation of the central uplift. The next phase was more widespread, folding the Paleogene flysch, that is, the rocks in the foredeeps on either side of the central uplift. This phase is associated in time, however, with the involvement of the flysch zone in the uplift. The later, Upper Tertiary folding phases follow; these were more strongly developed along the periphery of the folded zone, in the flanks of the expanding central uplift, and in the foredeeps. As far as the Pennine axial zone is concerned, a considerably less intensive formation of folds continued there until the pre-Pontian phase, after which the folding ceased altogether.

In the Upper Miocene subsidence took place in the depression composing the western Mediterranean, which in the terminology of this

book is an interior depression. It was superimposed on a considerable area of Alpine folding, specifically encompassing the coastal zones of the western Alps along their axes, in the zone of the central uplift. This caused a subsidence of this zone, so that it came to be filled with Upper Miocene and Pliocene sediments. These sediments lie undisturbed over the strongly folded older deposits, with a sharp angular unconformity, whereas at the margins of the folded zone, in the foredeeps, these younger sediments are folded to an equal degree.

Here it appears that, beginning with the inversion, the formation of folds not only spread over an ever greater area but also advanced toward the periphery of the folded zone while dying out in its center. In all other instances that may be cited here, only the main and terminal phases could be studied. The angular unconformities in the Caledonian folded structures of England indicate that the pre-Gotlandian phase was chiefly developed in their central uplifts and the later, pre-Devonian phases in the intermontane basins (Belousov and Gzovskiy, 1945). In the Pyrenees (the axial zone) the main phase is pre-Oligocene, whereas the folded structures on the periphery of this region have been produced by the Miocene and Pliocene phases (Misch, 1936).

A classic example of the migration of phases constantly mentioned in geologic works is the Hercynian folding in Western Europe. Since the Hercynian folded zone of Europe was overlapped to a considerable degree by the Alpine folding and was reworked in the Alpine cycle until it lost its identity, and since in other places it is concealed under a mantle of undisturbed Mesozoic and Cenozoic rocks in the depressions of the Alpine platform, the Hercynian structures in Europe can be reconstructed only in certain individual areas where the folded Hercynian basement crops out on the anteklises of the Alpine platform. The most thoroughly studied is the northern flank of this zone in the Bohemian Black Forest, the Vosges, the central plateau of France, the Harz region, the Rhine Mountains, southern England, and the Armorican massif (Stille, 1924; Kossmat, 1921; etc.). This flank comprises the northern half of the central uplift (the so-called Moldanubian zone), complicated by superimposed basins containing coal-bearing lacustrine deposits of Late Carboniferous and Permian age, as well as the foredeep and the marginal basin. It is established that in the south, in the Moldanubian zone (the Vosges, the Black Forest, and Bohemia) the folded structure was produced in the pre-Viséan phase. In the northern Vosges, the southern part of the Rhine Mountains, and the northern part of the Bohemian massif, a pre-Middle Carboniferous phase occurred, which probably affected the structure of the Moldanubian zone as well. Still farther north (in the Harz and Rhine Mountains), there is evidence of a pre-Permian phase. Finally, along the northern margin of the folded zone (in the foredeep), the folding was produced by the

Permian phases, which in the areas farther south were manifested by steep thrusts but which were totally absent in the Moldanubian zone. In the interior depressions of the latter zone the Upper Carboniferous is undisturbed, in contrast to Westphalia, where it is strongly folded (Fig. 212).

A symmetrical migration of the phases in the same folded zone, southward from the same central uplift, can be observed from the central plateau of France southward, toward the Pyrenees (Stille, 1934). The pre-Middle Carboniferous phase appeared along the southern edge of

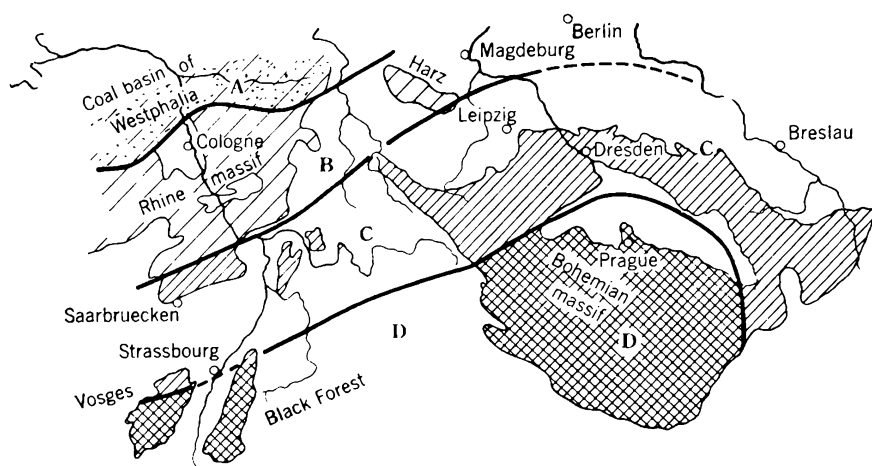


FIG. 212. Diagram of the migration of the folding phases in the Hercynian folded zone of Western Europe: (D) Danube Moldau zone (pre-Early Carboniferous and pre-Middle Carboniferous phases), (C) Saxon-Thuringian zone (pre-Middle Carboniferous phase), (B) Rhine-Hercynian zone (pre-Early Permian phase), (A) foredeep (pre-Late Permian and pre-Early Triassic phases).

the plateau, whereas the later phases developed in the Pyrenees, which correspond to the Hercynian intermontane basin.

MAIN FEATURES OF THE DEVELOPMENT OF FOLDING IN GEOSYNCLINES AGAINST THE BACKGROUND OF OSCILLATORY MOVEMENTS OF THE EARTH'S CRUST

It must be said at the outset that none of the phases of folding taking place in a geosyncline during a given cycle is manifested throughout all the geosyncline, but only in certain zones and frequently in very limited areas. Different phases are developed over different parts of the geosyncline; hence the migration of phases.

Furthermore, *there is a close association between folding and uplift of the earth's crust.* Before the inversion, the folding phases are con-

centrated in the intrageanticlines. After it, they take place in the central uplift; as the latter expands, the folding embraces a larger and larger area. When a central uplift contains an interior depression, the folding ceases in its axial portion but continues along its periphery, migrating away from the axis of the uplift.

The same relationship between folding and uplift may be expressed differently: *the intensity of the folding and its type are related to the intensity of the uplift of the earth's crust.* The intensity of the preinversion uplifts in the intrageanticlines cannot be compared with that of the postinversion central uplift: the latter is much more violent. For this reason, the folding that occurs before the inversion in the intrageanticlines is weak and of the intermediate or discontinuous types, whereas the postinversion folding is distinguished by its considerably greater intensity. The central uplifts do not reach the same magnitude everywhere: along their trend one encounters various degrees of uplift, and even of replacement of uplift by subsidence. Moreover, the amount of the uplift decreases away from the axis of the central uplift and toward the intermontane basins and the foredeep, and the folding accordingly decreases in intensity away from the interior of the central uplift, both toward its plunge and toward its flanks. There is a simultaneous change in the nature of the folding—from continuous to intermediate and discontinuous. Folds of the latter types are characteristic not only of intermontane basins and foredeeps but also of the plunges of anticlinoria (such as the Taman and Apsheron peninsulas of the Caucasus).

The relationship between folding and uplift of the earth's crust is also revealed in the fact that *each phase is accompanied by a separate brief uplift of the earth's crust in the place of folding.* Each of the preinversion phases is accompanied by a brief uplift within the intrageanticline. Since subsidence generally prevails in a geosyncline prior to the inversion, such uplifts, by opposing the general tendency, retard the development of folding and impart a zigzagging tendency to its course. Each such uplift causes erosion of a portion of the sediments and an unconformity below the deposits accumulated during the new subsidence.

The postinversion phases occur under the opposite conditions: those of rising uplifts. But even here each phase is accompanied by a brief and sharp uplift. In this case, the folding process reenforces the general tendency and accelerates the elevation, although after the brief upward push there is usually a reversal of movement—a subsidence and narrowing of the central uplift—after which its growth is resumed. Each of these brief upward movements, occurring against the background of a more general uplift, is recorded in erosion of the uplift and redeposition along the flanks of the central uplift.

The connection between the folding and uplift is so close that occasionally there is a local substitution of one for the other. Very often an intensive phase of movement, which is manifested in one place as folding, is represented only by uplifts without folding elsewhere within the same geosyncline. In such cases, erosion and redeposition are observed but no angular unconformity. For example, in the Greater Caucasus the brief uplifts—pre-Upper Liassic, pre-Callovian, pre-Tithonian, pre-Upper Cretaceous, and pre-Maykopian—are observed over greater areas than is the related folding. For instance, the pre-Upper Cretaceous uplift is very strongly manifested in the northwestern Caucasus, but with no accompanying folding.

Since the greater uplifts occur on the sites of great previous subsidence—the same being true of the smaller movements—the association between folding and oscillatory movements is frequently taken to be the reverse of the one described here: the contention is that the maximal (continuous) folding is developed in areas of the maximal previous subsidence. In this view, the greatest folding within a geosyncline is developed over the areas of most intensive subsidence, in the intra-geosynclines; conversely, the weakest folding is in the intrageanticlines. Both views are equally correct: continuous folding is associated with the areas of intensive uplift after the inversion in a geosyncline—that is, in the location of the intensive subsidence that preceded the inversion. What does matter is the succession of events: a great subsidence precedes the folding and a great uplift follows it. Thus it is correct to state that *the intensity of continuous folding is related to the intensity of the oscillatory movements*: it increases with the amplitude of the vertical movements.

This rule should be augmented by three more, all dealing with the relationship between folding and oscillatory movements.

The first is that *the horizontal movement of the rock masses determining the inclination of the folds (the inclination of their axial planes) is directed away from the regions of greater subsidence and toward the areas of less subsidence or of uplift, based on the distribution of the areas of subsidence or uplift before the inversion*. On the basis of the postinversion distribution, the relationships are naturally reversed: in other words, the movement of the masses producing the folds is directed away from the intra-geosynclines and toward the intrageanticlines. Inasmuch as after the inversion and the folding, an anticlinorium arises where there had previously been an intra-geosyncline and a synclinorium appears in the place of an earlier intrageanticline, a fan-shaped pattern of folding is observed, diverging in the anticlinoria and converging, or inclined inward, in the synclinoria, always provided that both are inverted. This order has been noted by many investigators and has been mentioned, in various forms, in geologic literature. Usually the move-

ment of the masses away from the interior of a geosyncline is believed to take the form of an overturning of the folds against the interior or exterior "rigid massifs," meaning the areas of thinner deposits and, accordingly, smaller amplitude of subsidence. In other cases, the older literature of tectonics mentions a tendency of the folds to be overturned in the direction of the depressions, referring here to the distribution of the areas of uplift and subsidence during the later stages in the history of the geosyncline—after the inversion.

Such instances can be multiplied. Let us consider the structure of the geosynclines whose history has already been discussed in this book. Figure 127 illustrates the relationship between the present folding and the distribution of intrageosynclines and intrageanticlines in the Celtiberian geosyncline. It can be seen that the folds in both the Guadarrama and the Demanda Mountains are inclined, fanwise, toward the Tagus and Ebro marginal basins, as well as toward the Douro interior depression. The history of these uplifts and depressions is a manifestation of the regularity stated above.

Figure 134 likewise compares the oscillatory movements and the folded structure of the Caledonian geosyncline in England. Here, too, one notes that the inclination and overturning of the folds are away from the intrageosynclines and toward the intrageanticlines.

A similar picture is observed in the Taurus-Caucasus geosyncline (Fig. 118), and a comparative scheme of the relationship between the inclinations of the folds and the oscillatory movements in the eastern Pyrenees (Fig. 213) has been presented by Aschauer (1936).

The connection between fan-shaped fold systems and anticlinoria is broken in areas characterized by parageosynclinal development and by the preservation of a noninverted uplift.

The second regularity is that *the folds are disposed parallel to the isopachs*. This explains many apparent inconsistencies in the orientation of folds. In the southeast of France, for instance, there is a small folded

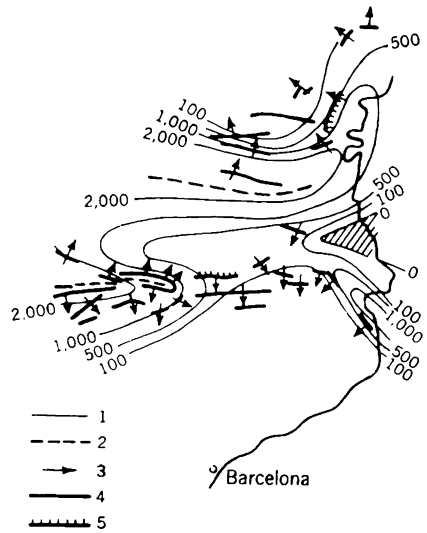


FIG. 213. Thicknesses of the Mesozoic deposits and inclinations of the folds in the eastern Pyrenees (after Aschauer, simplified): (1) isopachs of Mesozoic deposits, in meters; (2) axes of straight fan-shaped clusters of folds; (3) inclination of folds; (4) anticlines; (5) thrusts.

region, the Vauconte, which adjoins the Alps and appears to be a branch of them. But although the trend of the folds in this part of the Alps is parallel to the longitude, in the Vauconte system it is perpendicular, as illustrated in Fig. 214 (Richter, 1937). This is explained by the fact that the Vauconte folds were formed in an *embayment* of the Alpine geosyncline, whose axis and isopachs trend parallel to the latitude.

The Celtiberian geosyncline of Spain is a similar embayment of the Alpine geosyncline. Figure 126 shows how the trend of the folds exactly follows the outlines of the area of subsidence in this geosyncline.

The peculiar curvature on the map of one of the front ranges of the northern Caucasus, approaching the main cluster of Caucasus folds almost at a right angle in the vicinity of Ordzhonikidze, is also explained by the shape of the isopachs (Fig. 215).

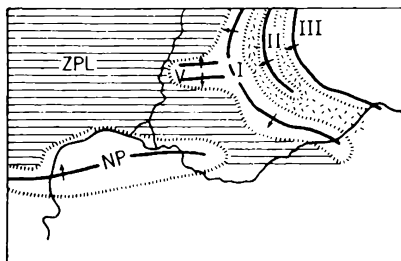


FIG. 214. Relationship between the folds in the Vauconte system and the Alps (after Richter): ZPL, central plateau of France; (I, II, III) Alpine anticlinoria formed in intrageosynclines; V, Vauconte folded system; NP, northern Pyrenees; the arrows show the direction of overturning of the folds.

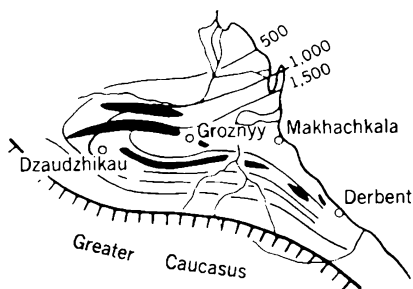


FIG. 215. Isopachs of the Maykopian deposits, in meters, and the anticlinal folds in the Terek depression (Caucasus). Solid black areas are anticlines.

The arclike curvature of the folds in the Urals, near the plateau of Ufa, becomes fully comprehensible when the distribution of the Lower and Upper Paleozoic isopachs is plotted (see Fig. 90). In all these cases, the thicknesses that reveal the areas of subsidence before the folding must be taken into account.

This rule also explains the transverse folds that sometimes appear at the plunges of anticlinoria. Usually they are younger folds developed after the formation of the anticlinorium in places where the subsidence continues. The isopachs of the sediments deposited here are, roughly, concentric around the anticlinorium, parallel to the long axis of the anticlinorium and perpendicular over its plunge. According to the rule of parallelism of folds and isopachs, the trend of the folds should follow the axis of the anticlinorium along its long sides, and it is quite possible for transverse folds to appear over its plunge. Such transverse folds, curved into arcs, are, for example, seen over the western plunge of the

western Sayan anticlinorium, corresponding to the closure of the western Sayan intrageosyncline (Kuznetsov, 1948).

The third and most important regularity is that *the intensity of the folding is determined by the thickness gradient*. The greater the gradient of thickening of the deposits, the greater the intensity of the folding. The present writer considers this rule to be of great theoretical importance.

Figure 216 shows a structural section through the Vauconte folded region and its relationship to the preinversion subsidence (Richter, 1937). This geosyncline is asymmetrical: its axis is displaced toward its southern margin, and thus the thickness gradient is greater on its southern flank. The corresponding folding is more intensive over the southern flank, and the inclination of the folds more pronounced than it is over the northern flank, where the folding is not only less intensive but is of the intermediate rather than continuous type.

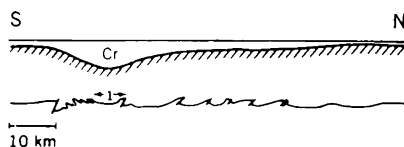


FIG. 216. Subsidence in the Vauconte region (above) and its present structure (below) (after Richter): *Upper diagram*, thickness of the Cretaceous deposits (vertical scale exaggerated five times); *lower diagram*, structure of the top of the Jurassic (vertical and horizontal scales equal).

The Welsh intrageosyncline of Great Britain is also asymmetrical: its northwestern flank is steeper than its southeastern. Correspondingly, the folding over the former is more intensive and tighter than over the latter, where the folds are intermediate rather than continuous.

The same regularity is clearly manifested in the structure of the Greater Caucasus. The great thicknesses of the Mesozoic deposits on the southern slopes of the main Caucasus range decrease rapidly southward, toward the Kura depression; at the same time the intensity of the folding along the southern slopes of the Caucasus is very considerable. It is noticeably less along the northern slopes, as is the inclination of the folds, which is again associated with the smaller thickness gradient.

If one considers this rule and generalizes from it, one will come to the conclusion that within the geosyncline the intensive folding is due not so much to the magnitude of the subsidence and the correspondingly great accumulation of sediments as to the fact that there is an alternation of sharply differing regimes of oscillatory movements within a small area, with a corresponding transition from great thicknesses to small, and vice versa, characterized by a high gradient of thicknesses. By the same token, the absence of continuous folds on a platform is

associated with small differences in thickness and small contrasts in the oscillatory movements. This conclusion enables one to understand certain phenomena that would otherwise be obscure. It is well known that the Lower Paleozoic of the Siberian platform reaches thicknesses up to several kilometers. Such thicknesses could be taken as typical of a geosyncline. On the other hand, no geosynclinal, or continuous, folds have been observed anywhere on the platform, which differs from a geosyncline not in the thickness of its deposits, but in their nongeosynclinal distribution: the thicknesses are more or less uniform throughout, and the thickness gradient is small.

Stille has expressed the idea that the weakening of a part of the earth's crust is caused by the subsidence of the crust into regions of high temperatures, thus allowing it to be folded. Hence he takes subsidence to be the "mobilizing" factor increasing the mobility of the earth's crust (Stille, 1924). It is evident from the foregoing, however, that subsidence alone is not enough. For continuous folds to be formed, there must be a considerable thickness gradient.

It may seem at first glance that this rule somewhat contradicts what has been said earlier regarding the connection between the intensity of the folding and the magnitude of the previous subsidence of the earth's crust and the succeeding uplift. Actually these are two manifestations of the same relationship between the oscillatory and folding movements. Great subsidence is accompanied by correspondingly great uplift, and vice versa. Thus the greater the magnitude of these two movements, the greater the contrast between them and, consequently, the greater the thickness gradients.

CHAPTER 23

THE ORIGIN OF CONTINUOUS FOLDING

DIRECTIONS OF THE FORCES IN FOLDING

The origin of the continuous folds in the bedded strata of the earth's crust has up to now been one of the most controversial problems of geotectonics. The first scientific attempts to solve the problem were made in the period when the contraction hypothesis was widely accepted; the most important works of this period are those of Suess (1875), Heim (1878), and Willis (1893). Later, after the contraction hypothesis had lost its leading position, the problem of folding was developed from similar views, as may be seen in works by Heim (1919), Sonder (1922), Stille (1924), and Kober (1926). Certain concepts of the contraction hypothesis as applied to folding continued to be elaborated even in more recent years, both in the Soviet Union and in other countries.

On the other hand, as early as the first decade of this century a growing number of authors pointed out some characteristic features of the structures and history of folded regions that were inconsistent with an association between folding and universal horizontal compression of the earth's crust. These authors considered the horizontal compression required to produce at least certain types of folding to be a secondary and local phenomenon derived from other processes.

At this point some mention must be made of the remarkable work by the Austrian geologist Ampferer (1906), which has not lost its significance even after fifty years. Ampferer's basic conception was that the horizontal compression that produced the folds in geosynclines arose within the geosynclines themselves and was in no way connected with any pressure from the platforms surrounding the geosynclines. Ampferer attributed this local compression of the earth's crust to the latter's reaction to the subcrustal flow of magma. This conception of the leading role of subcrustal currents, mainly convection currents, in the development of the earth's crust was later greatly elaborated in the field of geophysics. In its application to folding, this view was most concretely developed by Griggs (1939).

At the same time, another basic trend of ideas was developing. Its adherents saw the cause of folding as a local phenomenon in the stresses arising in the earth's crust from its vertical, wavelike oscillatory movements. In the past twenty years much ground has been gained by the hypothesis of gravitational folding, which associates the crumpling of the beds with their sliding (flow) down the slopes of tectonic uplifts, owing to the force of gravity. This hypothesis, initially proposed by Reyer (1888), was resurrected by Haarmann (1930) and elaborated with particular thoroughness by van Bemmelen (1933, 1955).

The particular approach to this problem by Soviet researchers in recent years has been to distinguish, among the various folded deformations, different types with different origins. The first and basic steps in this direction were made by M. M. Tetyayev (1934, 1941), who distinguished an extensive group of "domes" and attributed their formation directly to vertical movements of the earth's crust. From that time on, the notion that folded deformations had many different origins has been widely accepted among Soviet geologists. Following this conception, the present author (1947) distinguished idiomorphic (discontinuous) and holomorphic (continuous) folding, whose formation is due, respectively, to vertical movements of the earth's crust and to its horizontal compression. Many more types of folding have been described by V. Ye. Khain (1954).

In addition, Soviet investigators have pointed out a possible new mechanism that associates the folding caused by horizontal compression with wavelike oscillatory movements of the earth's crust. This mechanism will be called *dynamic squeezing*. This refers to the squeezing out of plastic beds from the crest of a growing tectonic uplift into its sides by the resistance exerted by the overlying beds against the growth of the uplift. It is suggested that the material in the plastic beds, in the crest of the uplift being formed at a certain depth below the earth's surface, is flattened and squeezed out to the side and, in the process, crumpled into folds. This idea, in its general form, was first proposed by M. M. Tetyayev (1941), and applied concretely to actual cases by the present author (1947, 1949), by M. V. Gzovskiy and A. V. Goryachev (1951), and by V. V. Bronguleyev (1951).

Analysis of the existing data leads to the categorical conclusion that the folding in geosynclines cannot have been produced by horizontal compression of the geosynclinal area as a result of pressure applied from outside. The following circumstances contradict the hypothesis of external horizontal compression:

1. *The shapes of the folded zones.* The Alpine folded zone of Europe has very complicated outlines, and the folds in it form a number of arcs: the Carpathians, the Alps, the Baetic Alps, the Apennines, and the Balkan Mountains. These arcs together form several oval regions of

folding, which are linked together in a chain and coincide exactly with the ovals of the Alpine geosyncline. One oval occupies the western part of the Mediterranean Sea, another is fringed by the Carpathians on one side and the Dinaric Alps on the other, and a third is formed by the Balkan Mountains and the mountains of Greece (see Fig. 113). This subdivision of the Alpine folded zone into ovals is observed not only in Europe but in other parts of the world as well; it is due to the fact that the mountain ranges formed by folds of Alpine age frequently form arcs.

In regard to the existence of external tangential forces, the problem may be stated as follows: in the rigid crust of the earth, which is incapable of being crumpled, there are ovals containing plastic material. These ovals are connected, forming the links of a single chain, but various parts of the chain have complex bends and blind ends, in which the terminal ovals are surrounded on three sides by rigid parts of the earth's crust. One such blind end is the westernmost Mediterranean oval, which in the area of the Straits of Gibraltar is surrounded on the north, west, and south by platform areas.

The Pyrenees are an isolated region of folding with two blind ends. The folds of the Donets basin die out to the northwest, grading into the platform. The Greater Caucasus, like the Pyrenees, is an isolated folded region bounded not only on the north and south but at both its ends as well by regions of discontinuous folding. Individual groups of folds on the western slopes of the southern Urals terminate in blind ends and die out en echelon along their trend.

All these facts quite thoroughly contradict the hypothesis of external tangential forces. Wherever the folded oval or group of folds terminates in a blind end, the area would have had to be subjected to pressure from three sides and this pressure from different directions would have had to originate in a single rigid body, which could be displaced only as a whole and only in any one direction.

2. *Transverse folds.* It was mentioned earlier that certain folded regions contain folds whose trend crosses the main trend of the folded zone. It is clear that tangential compression might produce folds with a single trend perpendicular to the direction of the pressure. Folds whose axes are parallel to the direction of the pressure under these conditions are mechanically impossible.

3. *The history of the folded zones.* The development of the folding in time within the intrageosyncline moves from the axis to the periphery; the earlier phases appear in the interior of the zone and the later phases primarily at its margins. Complex folded zones may contain several nuclei from which the folds spread, but from each such nucleus the folding moves out from the center to the periphery. This circumstance also contradicts the notion of external tangential forces; if the latter

were the cause of folding, their action should first involve the margins of the geosyncline and only then the interior.

On this basis, it would also be impossible to understand the great intensity of the folding in the center of the folded zone and its dying out toward the periphery, grading gradually into discontinuous folds, which are clearly the result of vertical and not of tangential forces. The migration of the foredeeps from the geosyncline to the platform is also inconceivable if the platforms are considered as rigid props.

All these features of the structure and development of folded zones quite clearly show that the folding does not result from external tangential compression and contraction of the original surface of the geosynclines. We are thus forced to conclude that the horizontal compression reflected in the crumpling of the beds into folds must have taken place not through the movement of rigid platforms, squeezing together the plastic geosynclinal zone between them, but as a result of processes occurring within the geosyncline itself. These processes displace the material in the geosyncline in such a manner that in certain zones there is tangential compression, with the formation of continuous folds.

Within the geosyncline there is an outward plastic flow of rocks from certain axes, and along these axes the mass of rocks is drawn off and the sedimentary strata decrease in thickness; in other places, where this flow encounters resistance and comes to a halt, the material is piled up and packed together and the beds may be crumpled into piled-up folds. Since the geosyncline may contain several such centers of outflow of the material and more than one or two directions of flow, the ovals, the blind ending of groups of folds, and the transverse folds are all fully possible as results of this mechanism, which must be fundamentally predetermined by the genetic connection between folding and the vertical movements of the earth's crust. Only if it is considered that the vertical movements taking place within the geosyncline in some manner produce the folding of the beds as a secondary process can one understand the various tectonic movements that develop within geosynclines and their basic regular features.

MECHANICS OF THE FORMATION OF CONTINUOUS FOLDS

The most difficult part of the conception that the movements involved in continuous folding are secondary to the oscillatory movements is the problem of how this process leads to the transformation of the vertical crustal movements into folding.

This transformation is associated with two processes: with the movement of the mass of the earth's crust along the slopes of tectonic uplifts under the force of gravity, and with the squeezing of the material in the beds out of the crests of the uplifts as the beds are flattened ver-

tically in the formation of the uplifts. Both these mechanisms lead to the same result: displacement of the material from the crest into the flanks or the neighboring depressions. In the course of this process, the total thickness of the strata becomes smaller at the crests and larger in the depressions. In bedded rocks, because of their mechanical properties, the increase in the total thickness takes the form of folding. Thus the folds are formed in areas where the displaced stratified material is piled together.

In interpreting the structure of folded regions from the standpoint of both these mechanisms—gravitational and dynamic squeezing of the material from the crests—one must obviously look for the zones of outflow from which the material was removed by the force of gravity or by squeezing and for the zones in which the material has been injected or piled together and accumulated in the form of folds. It seems most likely that the zones of outflow must be located in the axial parts of folded complexes and especially of central uplifts, on the slopes of which the most intensive folding is observed. It has been suggested that a single such zone of outflow “accommodates” the whole given folded complex and must thus extend over a considerable area.

Difficulties arise, however, in applying these assumptions to the actual structures of various folded zones, since there are no indications of the existence of extensive zones of outflow of the material in the axial parts of many folded regions. One such difficulty appears, for instance, in the southeastern Caucasus, inasmuch as its axial zone, from which the material should have been pressed out, is shown in the published geologic sections as a steep fan of strongly compressed and highly isoclinal folds; this, of course, contradicts the entire conception.

There are two possibilities: either the basic conception is in error, or else our knowledge of the structure of the internal parts of the folded zones is still insufficiently accurate. The second possibility is, at least, quite likely, since it is well known that the structures of zones with complex dislocations have been studied too schematically in general geologic surveys. Thus we must begin by trying to determine the probability of the second assumption.

From this standpoint, there are some interesting results of special detailed structural investigations completed recently to this end by a group of geologists, under the present author's direction, in the Caucasus, the Fergana region, Tadzhikistan, and the Kerch and Taman areas. This writer has also had the opportunity to make an expedition to the French Alps with the same goal. Through these investigations, it has been possible to discover a number of peculiarities of folded zones that up to the present have been either unknown or have attracted insufficient attention. The chief results of these studies are presented below.

It has turned out that in places of the most intensive oscillatory movements—that is, primarily in geosynclines—the uplift and subsidence are more or less stepwise in nature. Here the earth's crust appears to be divided by deep faults (Peyve, 1945) into separate parts that will henceforth be called *blocks*. These blocks are of varying width and are usually elongated along the trend of the geosynclinal zones of uplift and subsidence (sometimes they are oblique to this trend); they move up and down as entire blocks, within the framework of the subsidence or uplift, and are disposed as steps one above the other. The amplitude of these steps depends, naturally, on the steepness of the flanks of the uplifts and the width of the blocks. Their movement may be compared with the movement of small, immediately adjacent blocks of ice floating upon large waves in the ocean's surface.

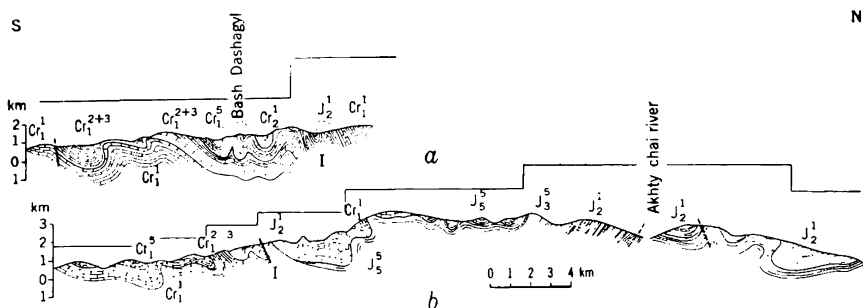


FIG. 217. Geologic sections through the southern slopes of the southeastern end of the main Caucasus range, in the region of Noukha (after A. Dolitskiy). The broken lines above the sections show the positions of the crustal blocks: (a) principal section; (b) detailed section through a part of section a south of the field I. One can see the crumpling of the strata immediately below the fault.

The following section will touch separately on the “primary divisibility of the earth's crust” and on its block structure. The blocks are separated by either faults or flexures, which commonly replace each other. Faults are undoubtedly the basic form of separation between blocks.

This steplike structure was noted some time ago in the case of the Tien Shan ranges, where the structural blocks were called *counters*. More recently such steplike structure has also been discovered in the Jura Mountains (Glangeaud, 1949–1950). There is also some basis for interpreting the structure of the high mountainous part of the French Alps as being composed of blocks (Beloussov, 1956). The discovery of such steps in the structure of the eastern part of the Greater Caucasus was completely unexpected. Figure 217 shows a section through the southern slopes of the southeastern Caucasus. In Fig. 218 are sections through the central part of the southeastern Caucasus (Shurygin, 1958).

All these sections clearly show steps, each of which, at the surface, is composed of rocks within a definite small stratigraphic interval. The surfaces of the steps contain folds, but the folds within each step are at a single level. In moving from one step to another, the age of the formations cropping out at the surface changes considerably. Similar steps in the structure of folded zones have been discovered in the northwestern Caucasus (Fig. 219) (Belousov, 1958). A study of the central part of the Greater Caucasus has shown that this high mountain

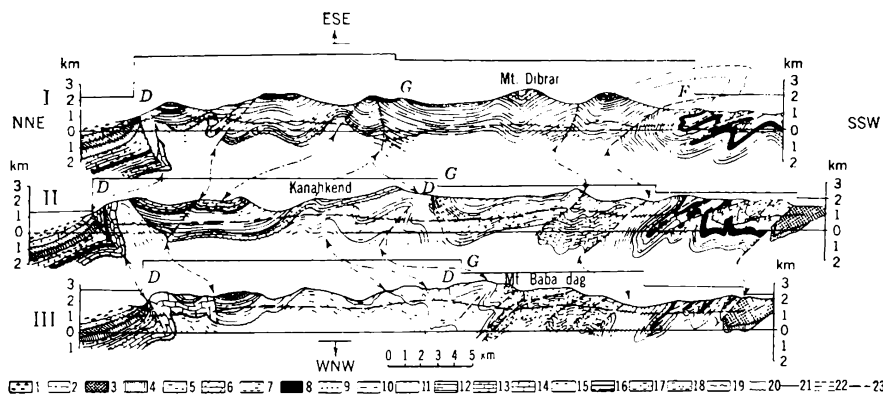


FIG. 218. Combined geologic profiles through the southeastern part of the main Caucasus ridge (between Mts. Dibrar and Bada-dag) (after A. Shurygin, 1958). From I to III the profiles are disposed from east to west. The broken lines above demonstrate the positions of blocks of the crust. (1) Upper Miocene; (2) Middle Miocene; (3) Oligocene and Lower Miocene; (4) Paleocene and Eocene; (5) Danian; (6) Upper Campanian and Maastrichtian; (7) Santonian and Lower Campanian; (8) Upper Turonian and Coniacian; (9) Cenomanian and Lower Turonian; (10) Albian; (11) Upper Aptian; (12) Hauterivian-Lower Aptian; (13) Valanginian; (14) Lusitanian-Tithonian; (15) Kimmeridgian-Tithonian; (16) Callovian-Lusitanian; (17) Middle Jurassic; points—sandstones of the Bayocian; (18) Lower Aalian; points—sandstones; (19) submarine slides; (20) stratigraphic contacts: (a) conformable, (b) disconformable; (21) tectonic faults; (22) lines indicating the continuation of structures: (a) of faults, (b) of anticline axis; (23) lower limit of the direct observations; D, injection folds, mentioned in the paper; F, southern overturned flexure; G, structural terrace, observed on the west and not visible on the east.

region consists of a whole “keyboard” of blocks that are elevated or depressed relative to each other (Fig. 220).

Observations in the Caucasus, as well as in other regions, lead to the conclusion that the type of folding that has been called continuous in this book is much less widely distributed in folded zones than was believed earlier, whereas the dislocations of strata that are immediately due to vertical movements of the crustal blocks are considerably more widespread. This mistake has resulted mainly from the fact that the different types of folding were not distinguished from each other and that in rapid surveys all the folded dislocations were grouped loosely

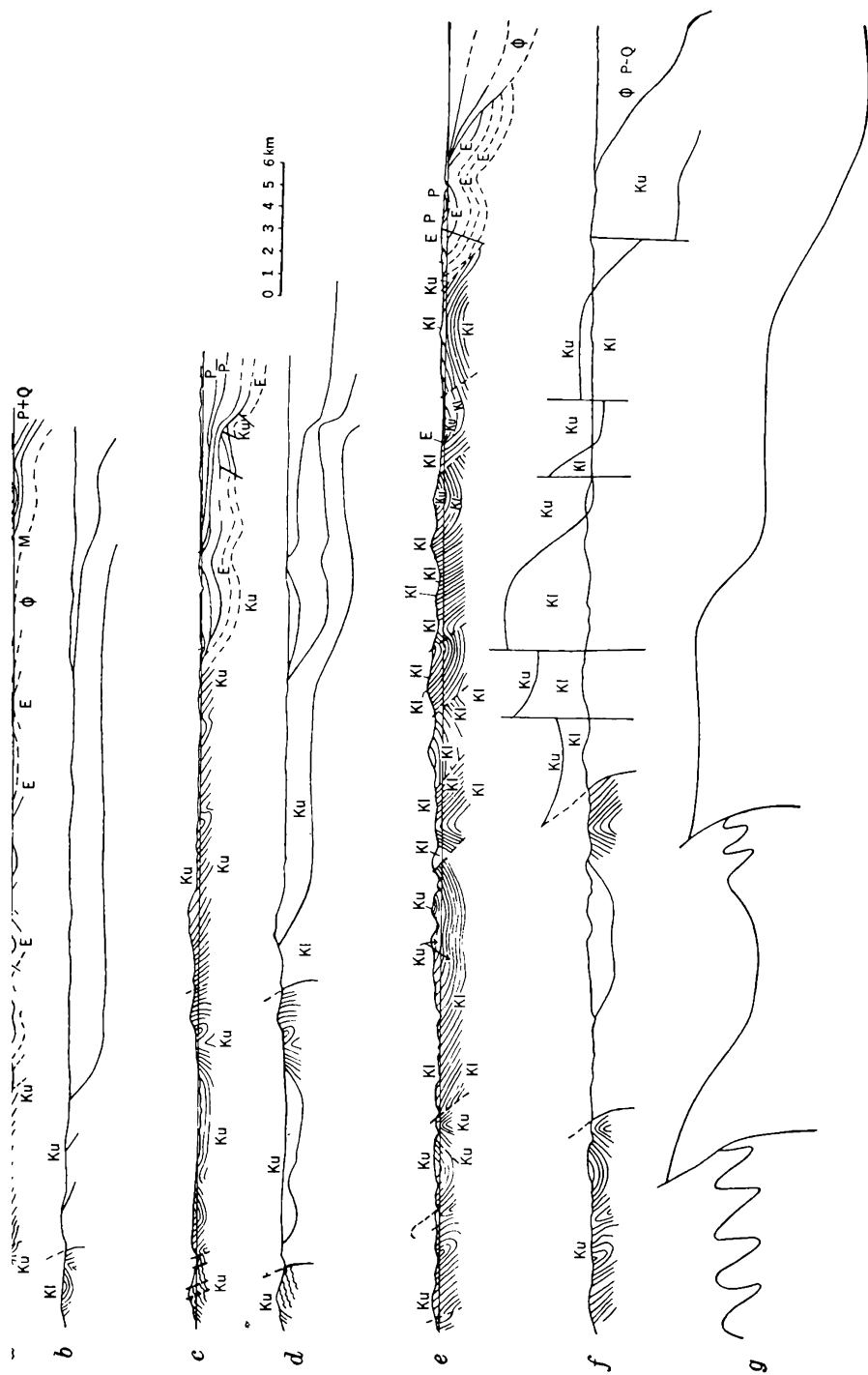


FIG. 219. Section through the northwestern Caucasus: (a, c, e) geologic sections (after various authors); (b, d, f, g) diagrams (after V. V. Belousov); (d) generalization of section a; (f and g) two stages in the generalization of section e.

under the category of "geosynclinal linear folding." A careful study of the morphology of folded complexes and a division of folds into their various types will radically change our ideas of both the structure of folded zones and the conditions under which folds are formed.

Moreover, even with a correct reconstruction of the morphology of the folds, there have been errors in their interpretation. All geologic

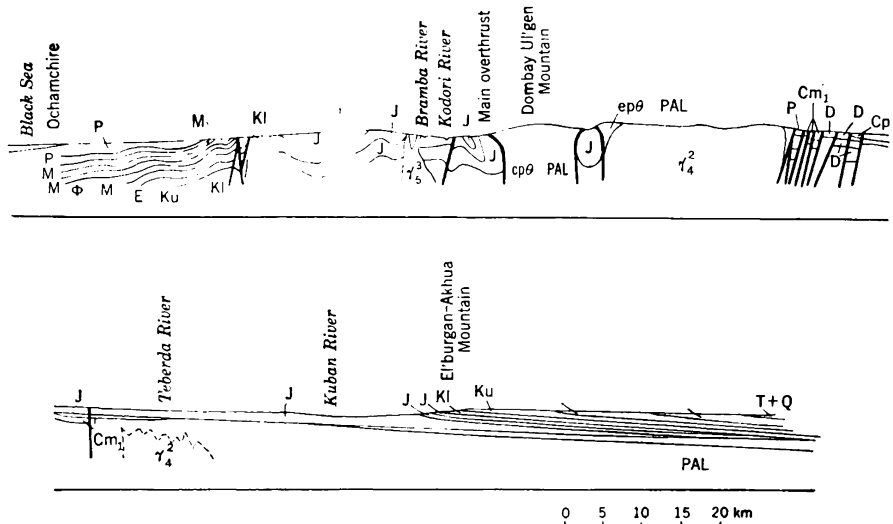


FIG. 220. Geologic section through the central Caucasus. (Top is left half of diagram; bottom is right half.)

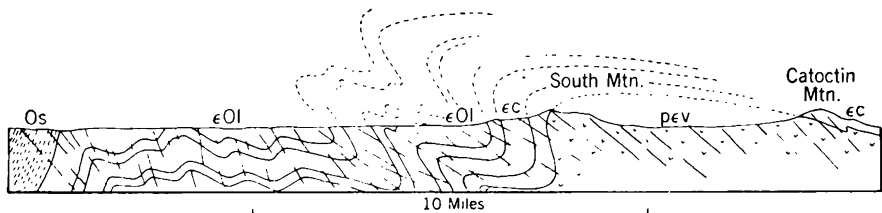


FIG. 221. Flexures in the western slopes of the Appalachians, which may be incorrectly interpreted as overturned folds (after King, from Eardley).

literature is filled with descriptions of "cascades" of overturned folds descending the flanks of anticlinoria to the bottom of the adjacent basins. But little attention has evidently been given to the fact that many such "cascades" of overturned folds are actually stairways of structural steps with almost horizontal surfaces, connected by vertical or slightly overturned flexures (Fig. 221). Thus these "cascades," once considered a fundamental proof of "external lateral pressure," have become one of the phenomena testifying to the block structure of folded zones.

The most general conclusion regarding the mechanism of continuous folding that can be drawn from the investigations described above is that the formation of continuous folds must be seen as the reaction of the beds in the earth's crust to the differential vertical movements of the blocks. Continuous folding is undoubtedly the result of a longitudinal bending of the beds arising from horizontal compression. But this horizontal compression is a strictly local phenomenon in the earth's crust and is due to local causes. Thus, for example, it arises in the vicinity of relatively uplifted blocks of the crust, in the rocks surrounded by these blocks. When a block of the earth's crust is elevated above the adjacent ones, there is a tendency to expansion or, more correctly, to lateral sliding in the upper parts of the uplifted block. If

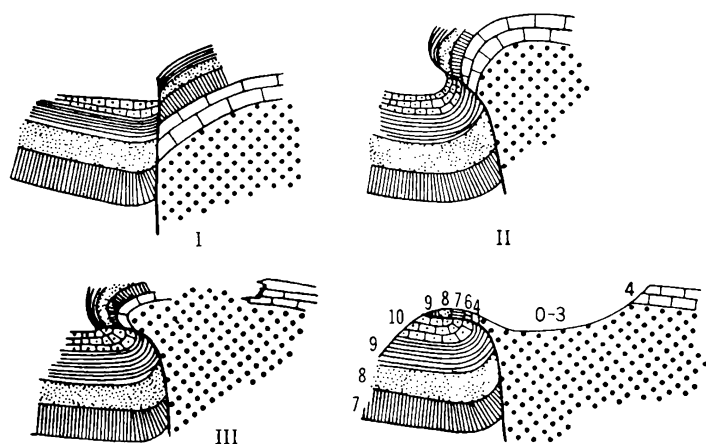


FIG. 222. Changes in inclination along the strike of a fault in the Jura Mountains: dots represent Triassic gypsum and salt; figures show the successive strata of the Jurassic (*after Glangeaud*).

the block is bordered by primary vertical faults, these faults are bent by this tendency in the upper layers of the block close to the earth's surface and the fault planes acquire an inclination; in this way, the vertical normal faults become upthrusts or overthrusts, and the block becomes fan-shaped. An example may be seen in Fig. 222, which shows the change in the inclination of the upper part of what had once been a vertical fault in the Jura Mountains. It is suggested that even the bending of the fault separating an uplifted block from the adjacent area, as in Fig. 223, is due to the same outward lateral sliding in the upper levels of the uplifted block. A no less clear picture can be obtained from Fig. 224, in which a single fault along its strike successively becomes a vertical fault, a steep upthrust, and an almost horizontal thrust.

This phenomenon appears very clearly in the sections mentioned

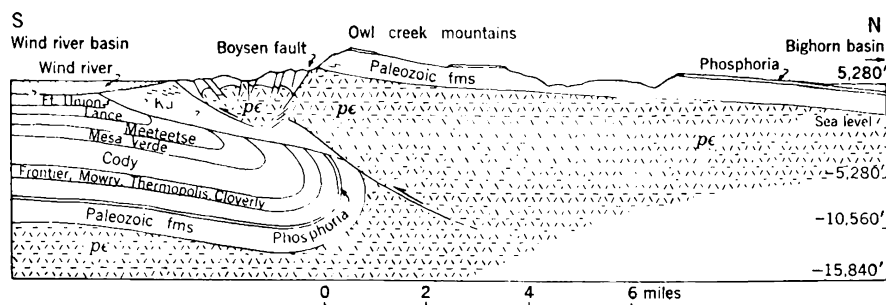


FIG. 223. Section through the Owl Creek Mountains (after Fanshawe, from Eardley).

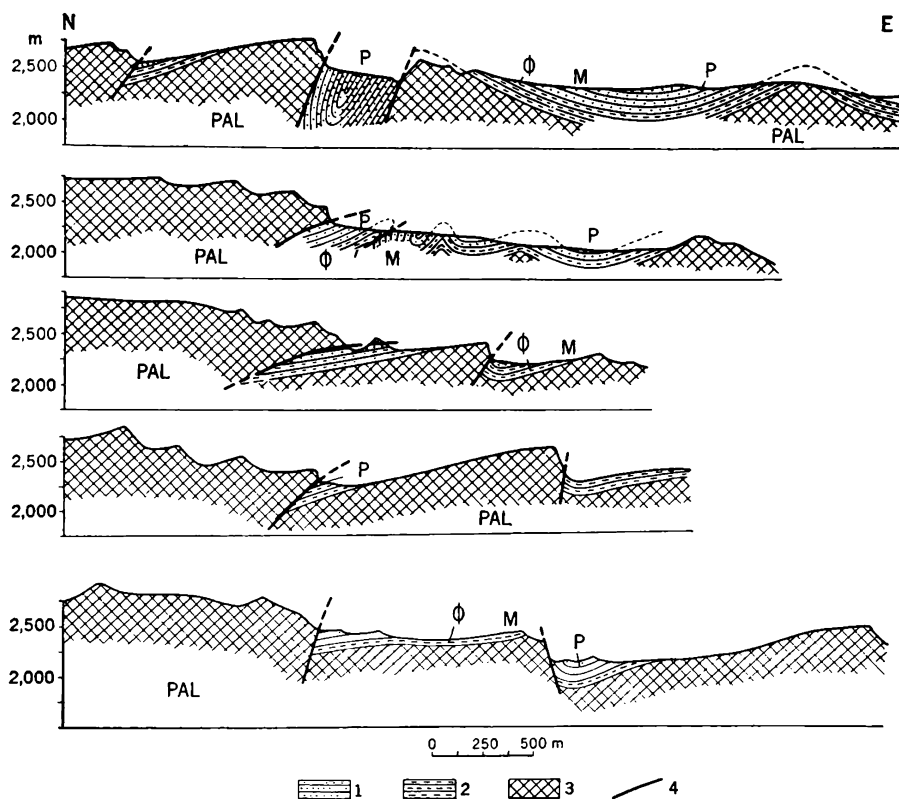


FIG. 224. Diagrammatic geologic section through the Kokdzharsu basin (in the Kirgiz Range of the northern Tien Shan) (after A. V. Goryachev): (1) Pliocene; (2) Miocene and Oligocene; (3) Paleozoic; (4) faults. From top to bottom, the sections follow each other successively from west to east. It can be seen clearly that the same fault (second from the left in the top section, left in the others) changes its inclination considerably along its strike.

above through the southeastern Caucasus (Fig. 218). In these sections, and especially in one of them (1F), there is a single uplifted block that has moved fanwise both to the south and somewhat to the north. This expansion in the upper part of the block presses the surrounding rocks horizontally, thus crumpling the beds in the adjoining regions into folds. The same thing may also be seen in Fig. 218 II. It is very clear in Fig. 219, which shows that in the northwestern Caucasus there is intensive folding of the beds under the overhanging edges of the elevated blocks. In the section through the central Caucasus (Fig. 220), each of the uplifted blocks of ancient rocks exerts pressure on the surrounding rocks and compresses them. There is particularly strong compression of the Jurassic rocks in a narrow depression between two nearby elevated

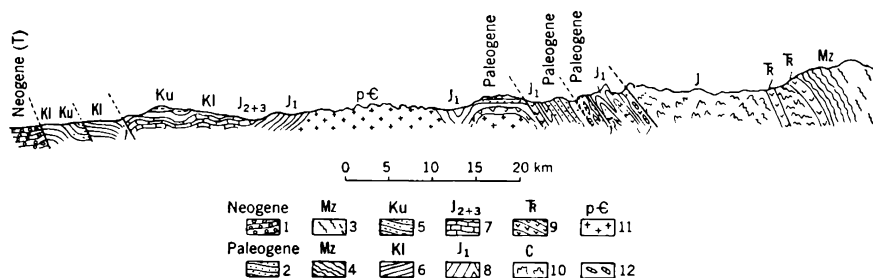


FIG. 225. Diagrammatic section through the Maurienne Alps, along the Aran River: (1) Neogene; (2) Paleogene (mostly flysch); (3) strongly deformed Mesozoic; (4) *schistes lustrés*; (5) Upper Cretaceous; (6) Lower Cretaceous; (7) Middle and Upper Jurassic; (8) Lower Jurassic; (9) gypsiferous Triassic; (10) Carboniferous and Permian; (11) crystalline rocks; (12) Mesozoic blocks tectonically thrust into other rocks.

blocks. Figure 224 also shows how the thrust of the Kirgiz Range has crumpled the Tertiary deposits of the adjacent depressions into folds.

If the blocks form a staircase, one higher than the next, as happens in the interior of folded zones, there is a unilateral thrusting of each higher block over each lower one. This is observed, for example, in the southern slopes of the southeastern Caucasus (Fig. 217). Here the blocks forming the steps are thrust over each other, successively compressing the beds near the surface into folds.

An extreme case of such a reaction of steplike blocks against each other is observed in the French Alps (Belousov, 1956). Geologists studying this mountain region have for many decades divided it into a number of longitudinal zones. These are separated by tectonic faults inclined at angles of 45° to 50° (Fig. 225). But all these zones (the Dauphiné, the Ultradapuiné, the Subbriançon and Briançon zones, and others) differ not only in their present structures but also in the sections of Mesozoic and Paleogene deposits composing them. The thicknesses and facies of these deposits differ in the different zones;

hence the zones must also have different histories of oscillatory movements. Like the keys on a keyboard instrument, they have moved repeatedly relative to each other in the vertical direction, sinking to different depths and being elevated to various heights. These differential vertical movements have evidently taken place along the faults separating the zones, so that initially these faults must have been vertical. They have probably remained vertical at depth, but the upper parts of the fault planes near the surface have become inclined. Thus the upper parts of the blocks appear to be overturned against each other, as in the Caucasus. This overturning in the Alps, however, is especially intensive, probably because the blocks here form steeper staircases. The upper parts of the higher blocks lie upon the upper parts of the lower ones and compress the beds in them. The beds of these compressed blocks are crumpled into tight isoclinal folds and broken into lenses. In the uppermost blocks occupying the axial parts of the range, which have not undergone compression from above, the beds are undisturbed and generally form a broad gentle syncline.

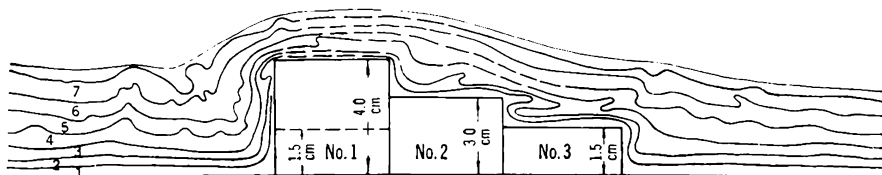


FIG. 226. Reproduction on models of gravity folding in blocks forming steps (after N. B. Lebedeva).

In the case of the Western Alps, a special role has been played by the extremely plastic Triassic deposits, which were very intensively compressed between two blocks and thus facilitated the high degree of deformation in the other rocks as well.

The conditions are almost the same when the blocks are separated not by faults but by plastic bends in the beds, or flexures. In that case, the expansion of the upper part of the block transforms the vertical flexure into an overturned fold, and beneath its overturned flank the deposits of the relatively depressed block are crumpled into folds (Fig. 218D and Fig. 219f).

It is interesting to compare these observations in the field with the results of experiments on small-scale models, particularly as depicted in Fig. 226. Here the bedded strata of the crust are represented by layers of resin. The folds that descend in a cascade down the steps of the uplifted synthetic blocks have been formed by gravitational flow of the resin from the higher to the lower steps.

There is scarcely any reason to doubt that in nature the main cause of the outward spread to the sides in the elevated blocks is the force

of gravity; what one sees is the result of plastic flow of the uplifted parts of the blocks caused by the weight of the rocks composing them. As in the experiment with the model, the rocks flow from the higher structural steps to the lower ones and in this process are crumpled into folds, especially directly under the vertical steps.

Another mechanism causing lateral expansion of the uplifted blocks is that of dynamic squeezing. Haller has produced a very interesting description of the structure of the "central metamorphic complex of northeastern Greenland." He indicated that the migmatites of the metamorphic basement were elevated into domes (the same as the blocks described in this text, but in a very plastic condition) and at the top flowed outward horizontally, under a thick cover of Paleozoic rocks, to form sheets or "mushroom" structures (Fig. 227). In this case, it is most natural to attribute the flow of the uplifted massif toward the sides to

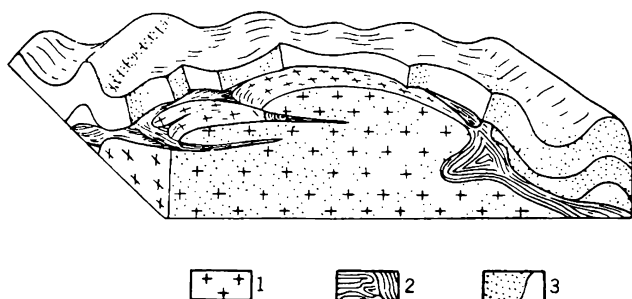


FIG. 227. Diagram of the structure of the metamorphic complex in Greenland (after Haller): (1) migmatites; (2) metamorphic slates; (3) Paleozoic sedimentary rocks.

resistance from the overlying rocks. The result of both processes—gravitation and squeezing—is the same. Dynamic squeezing must occur at some depth within the earth's crust, whereas the gravitational mechanism predominates close to the surface. But both these mechanisms, acting in parallel directions, may be closely intertwined.

Study of the structure of folded zones nevertheless leads one to conclude that not all the observed continuous folds have been formed by the means discussed immediately above. There are some folds whose origin would be difficult to connect with pressure from the adjacent uplifted blocks, inasmuch as they are removed to some distance from the latter or encountered on the uppermost of the uplifted blocks. In attempting to find an explanation for these folds, it may be noted that the discussion has thus far been limited to the mechanical aspect of the problem and has not touched at all upon the history of the folded structure. An examination of the development of the folds from the historical standpoint may provide additional clues in interpreting the mechanism of their formation.

Thus, for example, the uplift of the block may take place unevenly: first one side may be elevated and then the other. Moreover, at various stages the boundaries of the rising blocks may be in different positions, so that the places in which the folding occurs may also be displaced with time. But since the folds already formed are preserved, as the boundaries of the block are displaced the folds gradually cover a certain area. In section *F* of Fig. 218, the interrelationship between the beds shows that the northern (left) side of the block was elevated first, followed by the right side. In addition, if one compares the different sections in this illustration, one can see that the more western areas are broken into a greater number of blocks than the eastern: from west to east the blocks appear to be more and more combined. Although the easternmost section is essentially a single extensive uplifted block, folds are observed within it, whereas in the western blocks there is a structural step between blocks (Fig. 218*D*). Some folds may thus mark hidden boundaries between blocks.

A similar comparison of a number of sections through the north-western Caucasus reveals the same picture: toward the plunge of the Greater Caucasus anticlinorium, to the northwest, the division of the folded zone into blocks becomes more simplified, the number of blocks becomes smaller, and in the westernmost section one sees only one uplifted block (Fig. 219). But the surface of this single block contains a number of gentle folds which, as in the previous case, may mark latent boundaries between the blocks.

Another possible interpretation of the actual mechanics of formation of continuous folds is suggested by the fact that the blocks of the earth's crust repeatedly move upward and downward. This assumption, based on the oscillatory nature of the crustal movements, is well founded. Moreover, in experiments on models it has turned out that if one of the dies in the apparatus is pushed up higher than the next, and thus stretches the layers above it, and this die is then again lowered to its original position, the extended beds are too long and are thus folded as they settle into their previous position following the lowered die. By similar operations with several dies, one may obtain a model of an entire folded zone (Fig. 228). D. P. Rezva has used these repeated upward and downward movements of the blocks of the earth's crust to explain the formation, in certain box folds in the Mesozoic deposits of the Alayskiy Range, of additional marginal folds that rise like ears from the edges of the box folds (see Fig. 231*B*). It is suggested that in this case the block was first elevated above its present position, that the beds were elongated to accommodate this magnitude of uplift, and that the block was then somewhat lowered, so that the extra length of the beds was bent into folds.

A. V. Dolitskiy has suggested a more general form of the possible

role of this mechanism. The entire problem was given the concrete formulation it required after the experiments on models mentioned above, which were performed by N. B. Lebedeva. Thus there are now many possible explanations for the process of continuous folding as a result of the vertical movements of the earth's crust.

One more mention must here be made of the significance of the step-like structure of folded zones in the formation of folds. If there had been a gravity sliding of the material down the smooth flanks of uplifts, this would have led to the formation of folds mainly at the feet of the slopes, since without any resistance to the flow of material down the slope, the plastic rocks would have been piled up into folds only

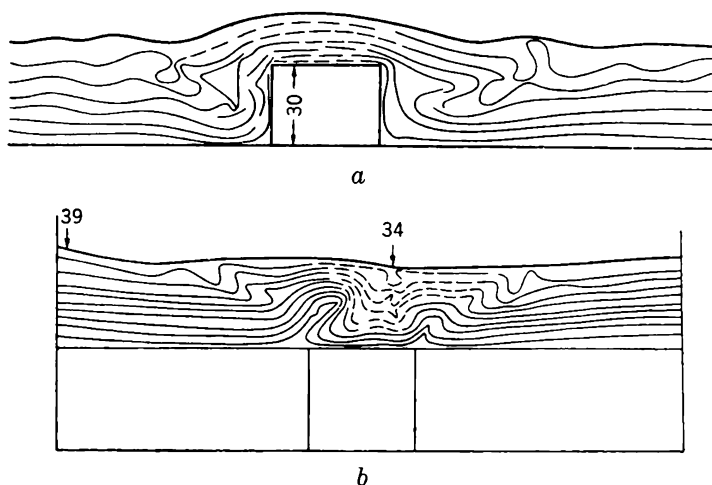


FIG. 228. Reproduction on a model of general compression folding by the successive elevation and lowering of a die; material is resin with smears of clay (*after N. B. Lebedeva*).

at the bottom. The steps constitute obstacles to this flow of material, acting as a succession of dams. But the folding actually covers the slopes of the uplifts more evenly and is not concentrated at the bottom alone. The boundaries between the blocks also control the trends of the folds, which are more stable than they would be in the case of a free flow of material down the whole slope.

These conceptions also provide an answer to the question of why it is impossible to find extensive zones in which the material has been displaced outward from the crests of the central uplifts. Local zones of outflow occur within each step, and each step also contains its zone in which the plastic material is piled up into folds. The whole folded zone consists of a series of paired adjacent local zones of outflow and folding. The distance to which the material is displaced in each case is thus not great.

CHAPTER 24

DYNAMIC CLASSIFICATION OF FOLDS. THE UNITY OF THE PROCESS OF FOLDING

DYNAMIC CLASSIFICATION OF FOLDS

The types of folding that have been distinguished—discontinuous, intermediate, and continuous—are morphologic varieties of folds. But the material presented above also provides the basis for distinguishing types of folds that differ in the method, or mechanism, of their formation. Since the mechanism of folding is, actually, the movement of the earth's crust or the rocks composing it, which is the direct cause of the formation of folds, this discussion must obviously deal with the dynamic classification of folding.

A bed may be bent into folds in two ways: transversely and longitudinally, with respect to the fold. In the first case, the bed is bent by forces normal to it, applied in a direction perpendicular to the bed. Since the bed is initially horizontal, the forces that produce transverse bending are directed vertically. If they act upward from below, the bed is uplifted and bent into an anticlinal fold. In the case of longitudinal bending, the bed is subjected to the action of forces in its own plane; these forces are horizontal. Thus the bed is bent into folds as a result of compression. The shapes and sizes of the anticlines and synclines and their disposition will be determined in this case chiefly by the reaction of the bed to this mechanical action—that is, by its mechanical properties—whereas in transverse bending the most important factors are the magnitude and location of the vertical forces. The main examples of beds bent transversely are discontinuous folds, as well as box folds in the intermediate category. This refers to the majority of such folds, although in particular cases another origin may be suggested.

It was shown earlier that discontinuous and box-shaped intermediate folds typically arise as the result of tectonic forces directed upward and manifested locally. The folds formed in this manner may be called *block folds*. This term can scarcely arouse any doubt in the case of box folds, which are bordered by faults or steep flexures. It may, however,

cause some confusion in regard to gentle discontinuous folds such as domes and swells. It was seen above, however, that in a series of plastic bedded rocks the bending of the beds must level out upward in the section, in such a manner that a sharp box-shaped uplift at depth must be reflected in a gentle bending of the beds near the surface. For such gentle folds, which are the surface expression of block movements at depth, the term *reflected block folds* may be used.

It must be acknowledged, however, that a box shape does not always indicate a block-movement origin of the fold. For example, if an injection core (see below) is overlain by massive limestones, observations show that boxlike uplifts are often formed in the latter; these box-shaped uplifts are apparently associated with the mechanical properties of the limestones.

Continuous folds are the result of horizontal crumpling common to a greater or smaller group of beds. These undergo longitudinal compression. From the standpoint of its mechanism, such folding may be called *general compression folding*, and one may distinguish a corresponding dynamic type of fold. In most cases, of course, this will be the same as continuous folding in the morphologic sense.

Still one more dynamic type of folding is distinguished, which may be called *injection folding*. This originates in displacement along the bedding of one very plastic suite of beds occurring between other suites; the plastic beds flow out from some places and are injected into others. The material accumulates in the latter, the thickness of the plastic suite increases, and the overlying rocks are uplifted, forming an anticline. This category includes diapir domes and folds, as well as the ridge-shaped variety of intermediate folds, which earlier in this book were also attributed to the same mechanism of subterranean flow of plastic material. This type of folding is thus characterized by division of the beds into an active, or mobile, complex and a passive complex. The former is sandwiched in between the latter, and the movement occurring in the active complex affects the structure of the overlying rocks. The lower passive complex plays no direct part in injection folding.

Forces of longitudinal compression act within the active complex where the material is injected. Actually, in the places where the injection occurs (the diapir cores), the beds of the active series are frequently crumpled into compressed folds. The active series of very plastic rocks consists of salt or gypsum or very soft clay. On a small scale, injection folds are also formed in thick coal beds, since under tectonic conditions coal may behave plastically.

The bending of the upper complex of rocks occurs under the influence of pressure from below, from the injected material of the active complex. The upper complex is subjected to the mechanism of transverse

bending. Thus it can be said that injection folding combines elements of the mechanisms of both overall compression (longitudinal bending of the plastic series in the points of injection) and block movement (transverse bending of the overlying rocks).

One may differentiate injection cores in which the plastic material elevates the overlying beds, but does not penetrate them, from piercement cores (or diapir cores), in which the plastic material pierces the strata above. In addition, piercement cores may be covered and occur in the depths (covered diapir domes) or squeezed out on the surface (open diapir domes).

The bottom of the active plastic series is a surface of disharmony; below it, in the lower complex, the beds are either undisturbed or else dislocated otherwise than in the upper complex.

According to the dynamic conditions of their formation, block folds are a uniform group of dislocations. They are all closely connected with vertical movements of blocks of the earth's crust and differ only in dimension, amplitude, and shape. These differences are due to differences in the intensity of the vertical forces, their horizontal distribution, and the thickness of the plastic layered rocks that undergo the deformation.

The matter of injection folds is somewhat more complicated. These may be divided, on the one hand, into injection folds of gravitational origin and, on the other hand, into injection folds associated with outward pressure from the crests.

Gravitational injection folds are formed by a simple floating or buoying of lighter material within heavier (usually diapir domes) or by squeezing of the plastic material by the action of an uneven load above; from the standpoint of physics, these are the same mechanism. But the uneven distribution of the load may be due to differential vertical movements of blocks of the crust, accompanied by an accumulation of sediments in the lowered synclines and erosion of the uplifted anticlines, or to irregularities in the relief, or, finally, to a combination of both these factors. If, however, the inversion of the relief is not on a sufficient scale, injection folds can be formed only when the plastic complex has a lower specific weight than the overlying rocks.

Actually, for a movement of the plastic material to begin, the downward-acting load on the horizontal (level) surfaces passing through the plastic series must vary in different places. This can come about through irregularity in the relief of the earth's surface or through differences between the specific weights of the rocks in the plastic complex and in the overlying beds. In the latter case, if the active complex is to be squeezed upward, its specific weight must be less than that of the rocks above it. Under these conditions, if an anticline is formed and eroded in the course of its formation, the pressure on a certain horizontal surface within the plastic series will be smaller, horizontal stresses

will arise in the material, and if these stresses are greater than the plastic limit, the material will begin to be injected into the core of the anticline (Fig. 229). One can easily see that if the plastic rocks are not lighter than the overlying rocks, such squeezing cannot take place. It can then occur only as a result of irregularity of the surface relief.

For the diapir cores to float up toward the surface, there must be some initial disturbance in the surface of the salt or gypsum to initiate the movement. Since, when heavier rocks overlie lighter ones, the mechanical equilibrium between them is unstable, such a disturbance can arise very easily.

Injection folds formed in the process of squeezing of rock from the crests are associated with essentially different circumstances. Here the movement of the plastic material is caused directly by differences in

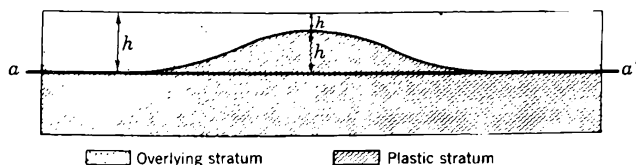


FIG. 229. Diagram of the formation of injection folds under the conditions of an uneven load applied from above. The density of the overlying series is considered to be 2.30; that of the squeezed plastic rocks is 2.15–2.25.

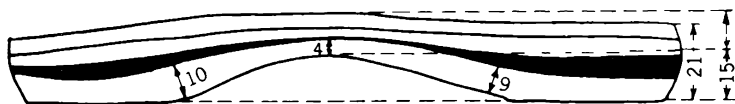


FIG. 230. Reproduction on a model of the formation of a dome in layers of wax; dimensions shown are in millimeters.

the pressure on this material. But these differences are produced not by the differing weight of the overlying rocks but by uneven pressure from below exerted against the resistance of the rocks above. The pressure from below arises from the rising folds originating in the movement of blocks. This pressure is local, and the plastic material is squeezed laterally into places where such pressure is less or nonexistent.

This process of squeezing may be reproduced on laboratory models. If one takes a model of four layers of wax (Fig. 230) covered by a layer of sand and then attempts to bend the wax layers upward, for example, by the expansion of a core of resin beneath them, instead of bending one will observe a sharp decrease in the thickness of the wax layers above the crest of the dome and an increase in their thickness in the flanks. Here the effect of squeezing is obvious.

General compression folding, we have seen, is produced by horizontal compression of local and varying origin. The mechanisms of gravitation and pressure again operate here. The horizontal compression arises

at the edge of the relatively uplifted block, which has either pushed its upper layers to the side under the force of gravity or else become inclined as a unit and leaned its weight against the adjacent block. Local compression arises as a result of gravity sliding of the rock from the higher levels to the lower. The folds may be formed in successive upward and downward movements of the block, which first stretch out the beds and then force them to reoccupy their previous area, for which they are now too long.

Morphologically, squeezing leads to the same results as the previous gravitational processes: to horizontal extension (and vertical contraction) of the beds in the uplifted areas and to their being squeezed out into the adjacent areas, where horizontal compression is created as a result of the squeezing.

UNITY OF THE PROCESS OF FOLDING

If one sums up everything that has been said earlier in this book about folding, one can perceive the single general basis for all the numerous mechanisms in the formation of the various types of folds. All folding is actually associated with differential vertical movements of blocks of the earth's crust. The most general conclusion is that the crumpling of beds into folds of any type is the reaction of the bedded strata of the crust to the vertical movement of the blocks. This reaction, in different cases, takes the form of different deformations of varying complexity. Such cases may be listed in order, from the simplest to the most complicated association between the block movement and the folding.

Block folding directly reflects the vertical movements of the blocks of the crust. The uplift of the latter changes, in reflected block folds, toward a leveling out of the uplift in the upper layers of the thick plastic series. The most typical manifestation of injection folding is due to the circumstance that the vertical movement of the blocks is complicated by horizontal flow in the most plastic series; this is caused by these movements of blocks in combination with the relationship between the specific weights of the rocks and the irregularity of the surface relief. The most extreme cases, which are offshoots of this general series of folded deformations, are those in which the above factors—the relief and the difference in density—act independently, apart from the displacement of the blocks, when the latter is either very small or totally absent (as in the floating of diapir cores and squeezing of the plastic rocks at the bottom of an erosional basin). Finally, general compression folding is a still further complication of the vertical movements of blocks, which are now accompanied by horizontal compression and crumpling of a thick series of rocks.

In considering the geologic conditions under which one type of folding or another appears, one must conclude that block folding, from

what has been said above, is the basis of the other types of folding as well and is best manifested in its pure form in places where the sedimentary mantle is relatively thin and the vertical movements of the crustal blocks are on a fairly small scale. These are the conditions on platforms, in certain parts of foredeeps and intermontane basins with a moderate accumulation of deposits, and over intrageanticlines.

Injection folding requires a thick sedimentary series containing a complex of plastic rocks and a moderate, "average" amplitude in the movements of the crustal blocks. The most favorable regions are foredeeps and intermontane basins with thick accumulations of deposits, deep synclises, and the plunges of anticlinoria.

General compression folding is associated with high contrast in the movements of the blocks, taking place with great amplitude and repeated changes in the direction of movement. Such conditions are typical

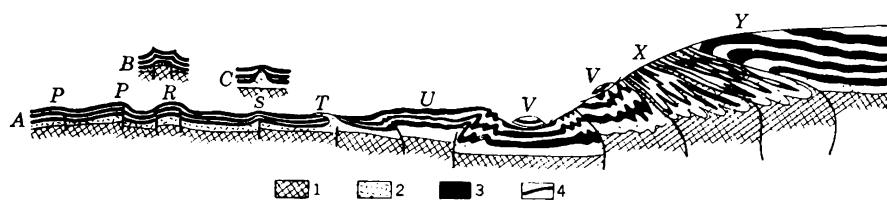


FIG. 231. Diagram of the structure of a folded zone: A, principal diagram, (1) crystalline basement, (2) highly plastic rocks, (3) different sedimentary rocks, (4) tectonic faults; B, box-shaped fold with two minor border folds; C, diapir dome; P, unilateral; R, bilateral block folds; S, injection fold; T, injection fold with piercement core; U, block fold bordered by overturned flexures; V, gravitational thrust; X, zone of crushed rock; Y, middle block, slightly deformed.

of geosynclines and also of intrageanticlines, in places where they subside most intensively.

It must be stressed once more that block structures arise in all the cases enumerated; although they appear in pure form only in the case of block folding, in two other cases they are to some degree complicated and camouflaged by the horizontal flow of the material in the earth's crust, merely "shining through" the structural forms of a different nature. Study of the concrete causes of folding in each case requires the investigator to discover the universal "block movement foundation." The present writer, for example, has attempted to do this in the sections illustrated in Figs. 217 to 221.

Figure 231 shows a diagram of the different types of folding and their interrelationships.

GENERAL REMARKS

It has been concluded that, although the principal basis of folding of all types is the same—the vertical movement of blocks of the earth's

crust—the immediate and concrete causes of folding vary greatly. They may differ even within individual, fairly small folded regions.

Up to the present time, the geologist, in studying the structure of a folded region, has usually not found it necessary to consider the causes of the folding (at least not thoroughly). These causes were thought to be so obvious and so universal (contraction of the entire surface of the earth!) that it appeared senseless to study them in each individual case. But the concepts described in this book make new demands on the geologist. They compel him, in any regional investigation, to study the means by which all the observed folds were formed. The geologist must ask himself: what are the *local* causes of folding, as revealed in the concrete local conditions of tectonic development? This, in turn, requires a completely new approach to the study of folding, which may perhaps be more laborious but is also immeasurably more specific.

It must also be admitted that there is a common aspect in all the varieties of concrete local mechanisms connecting the movements of blocks of the crust with the formation of continuous and injection folds. In every case, folds of this type are formed as a result of redistribution of material flowing from some places and accumulating in others. The flow of material is associated with a decrease in the thickness both of the entire folded series and of individual beds and with a corresponding increase in the surfaces of the latter. The compression that complements this extension of the beds occurs in places where the material accumulates. But since this compensation takes place not by a reverse increase in the thickness of the beds and a decrease in their surfaces, but through the crumpling of the beds into folds, the folds are due, in general, to the increase in the surfaces of the beds, which in being extended are forced to occupy the same area as before.

This conclusion is essential. Up to this time, the most widely held views were those based on the opposite assumptions: they regarded contraction of the overall surface of a given part of the earth's crust, along with retention of the surface areas of the beds, which were then forced to become folded in order to occupy this contracted space, as the principal cause of folding. The conceptions in this book are exactly the opposite. Efforts must be made to search for possible causes of the increase in the surface area of beds as the fundamental link in the mechanism of folding.

Some additional general remarks are in order. It must not be thought that the squeezing mechanism or gravitational mechanism of folding is limited to deformation of the nonmetamorphosed rocks of the sedimentary mantle. Many deformations also involve the rocks of the crystalline and metamorphic basement. Such involvement, for example, does not occur in the Jura Mountains, where the folded deformation is concentrated in the mantle of sedimentary rocks. But in the interaction of

adjacent blocks, as in the interior zones of the Alps, the rocks of the basement complex are subjected, along with the sedimentary rocks, to crumpling and deformation. Hence the commonly accepted idea of the "rigidity" of the crystalline basement must be radically reexamined. Metamorphic rocks, with their schistosity and their altered internal structure, are characterized by the oriented disposition of their minerals and by the development of minerals with flat and smooth surfaces; in many cases they undoubtedly possess greater plasticity than many sedimentary rocks under the same conditions. Accordingly, the crystalline basement does not react to the formation of folds of block origin only by rupturing; it may also be bent plastically, forming swells and domical uplifts.

The development of oscillatory movements and the association between continuous folding and the regime of oscillatory movements have been considered above. The concept of the block structure of the crust does not in the least alter the conclusions drawn earlier in regard to the succession and the laws of wavelike oscillatory movements. The formation of intrageosynclines and intrageanticlines, central uplifts, intermontane basins and foredeeps, etc., is subordinated to the same laws, whether the earth's crust is represented as being cohesive and bending plastically or as brittle and fractured. In the latter case, the crustal movements will be characterized by both blocks and waves (Khain, 1958) and will be discrete, as reflected in the blocks forming steps on the flanks of uplifts and depressions.

The block structure of the earth's crust also appears in the fact that the expansion and contraction, or the migration, of the zones of uplift and subsidence and the displacement of the boundaries of uplifts and depressions occur not continuously but by pulses, which draw an entire zone of the earth's crust into the uplift or subsidence. This peculiarity in the development of wavelike oscillatory crustal movements was pointed out by M. V. Gzovskiy (1948), who also spoke of the "keys" or "elementary zones" of geosynclines.

The parallelism between continuous folds and isopachs is evidently due to the fact that the blocks and the boundaries between them are predominantly elongated parallel to the contours of the intrageanticlines; by their own movements during the period of subsidence they determine the distribution of thicknesses. In certain cases, it has been noted that there are transverse steps across the plunges of inverted anticlinoria, as in the southeastern and northwestern Caucasus, where the Greater Caucasus anticlinorium plunges at both ends. Such transverse steps can clearly also produce transverse folds.

The connection between the intensity of the folding and the gradient of thicknesses is also easy to explain. Since the postinversion uplift is to some degree a mirror reflection of the preinversion subsidence, wher-

ever the intrageosyncline had a steeper slope, a correspondingly steep slope appeared on the postinversion central uplift, complicated by steep ladderlike steps on which there was an intensive formation of general compression folds. On the more gently sloping flank of the central uplift, corresponding in position to the gentler slope of the intrageosyncline, the staircase of blocks slopes less steeply, the steps are lower, and the continuous folds therefore are gentler or appear only locally if at all. These are the conditions in the main Caucasus range, for example, where intensive continuous folding is observed only on the steep southern slopes and is completely absent over the large areas of the gentler northern slopes.

This concept of the origin of folding makes it possible to understand why, in passing from a zone of greater to one of less tectonic disturbance, the continuous folds not only become more gentle but also grade into injection or block folds. The three types of folding may replace each other, depending on the intensity of the vertical movements of the blocks.

The epochs of continuous folding are associated in time with the appearance and growth of the central uplifts in geosynclines. The formation of these central uplifts creates the conditions for gravitational displacements and for dynamic pressing of the plastic material. If continuous—that is, general compression—folds develop, as indicated by observations, in separate thrusts or phases, these phases must be considered as the reflection of a corresponding discontinuity in the relative movements of the blocks forming the central uplift.

The vertical movements of the earth's crust are, generally, discontinuous in time. This was pointed out earlier in considering the development of oscillatory movements. As a result, the phases of folding coincide with the epochs of temporary increases in the contrast of the vertical movements of the blocks, the relative height of the steps, and the steepness of the slopes. During this time there is an intensified displacement of material by the force of gravity or by dynamic pressure and squeezing. The displacement stops when the gravitational leveling is sufficient and there is equilibrium between the force of gravity and the plastic limit of the rocks or when the contrasts and deviations are compensated by temporary reverse movements.

The differences in the development of continuous folding (by phases), on the one hand, and injection and block folding (long-lasting but uneven growth), on the other, are perhaps due to the fact that the intensive vertical movements of the blocks reveal the discontinuity of these movements more sharply than movements that are slower or of smaller amplitude.

Since folding is associated with redistribution of the material, the primary thicknesses of folded series cannot be preserved. This is the case

in zones where this material is squeezed out, where the decrease in the primary thickness of the beds is in certain cases very considerable. The present thicknesses of the sedimentary rocks in such zones cannot be used to reconstruct the history of oscillatory movements; the researcher, in studying all the phenomena of dislocation, must attempt to find a means of making the required quantitative corrections.

In zones of predominantly injection folding, the situation is otherwise, inasmuch as the increase in the total thickness of the rocks in this zone does not occur by an increase in the thickness of each bed separately but by compression of the beds into folds. The thickness of the beds themselves should not change significantly, apart from the fluctuations in the thickness of the beds in individual folds between the flanks and the crests and troughs.

Since squeezing out and injection are closely interspaced in folded zones and since zones of one or the other alternate on the slopes of central uplifts at short distances from each other and are grouped in pairs over each block, for a general analysis of the thicknesses that does not presuppose any great detail it is sufficient to use the average thicknesses for several blocks. A more detailed analysis of the thicknesses in folded regions must be based on a more profound study of the folded structure. It must be stressed, however, that in geosynclines the fluctuations in the thickness are so much greater than on platforms that under all conditions a quantitative solution to the problem of the disposition of zones of greater and less subsidence requires the use of thicknesses measured directly.

PART SIX

**TECTONIC MOVEMENTS
PRODUCING RUPTURE**

CHAPTER 25

SOME ASPECTS OF THE PHYSICAL THEORY OF FAILURE IN SOLID BODIES

Under geologic conditions, tectonic dislocations result from the failure of solid bodies under the mechanical action of forces applied externally. An attempt to explain the mechanism by which tectonic dislocations are formed will therefore make the greatest possible use of the modern theory of failure in solid bodies.

As in the case of plastic deformation, up to the present time no general theory has been elaborated. It has been possible only to make a certain summary of the existing empirical data and to draw some particular conclusions of a specialized character, which apply mainly to metals. Some of these data, however, may be used for our purpose, since the general features occurring regularly in failure seem to be the same for the most varied solid bodies (Uzhik, 1950; Fridman, 1946).

Experience has shown that every process of deformation ends in failure of the body, when the stress has reached the magnitude of the ultimate strength of the body under the given conditions. There are two types of failure: *parting* and *shearing*, or *shear*. Parting is pro-

duced by normal tensile stresses. It occurs when these stresses reach a certain critical magnitude and is reflected in the formation of fractures perpendicular to the principal axis of tension. A cylindric isotropic body that is subjected to tension along its axis and fails by parting will fracture in the direction perpendicular to its axis (Fig. 232a).

Shearing, or shear, is produced by tangential stresses and takes the

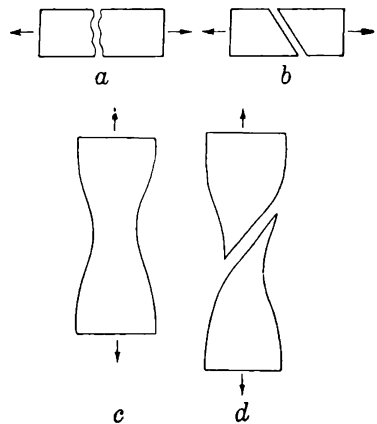


FIG. 232. Parting (a), brittle shearing (b), necking (c), and plastic shearing (d) in tension.

form of fracturing in the direction of the greatest shear. The same cylindrical body, subjected to axial tension and failing through shear, produces fractures cutting the cylinder primarily at an angle of 45° to the long axis (Fig. 232b).

Since plastic deformation is produced not by normal but by tangential stresses, parting is not directly associated with plastic deformation. It may, and frequently does, occur immediately after elastic deformation, as soon as the normal tensile stresses reach the ultimate strength. It may also set in after a certain amount of plastic deformation, as the stresses increase with the strain hardening of the body. But here, too, the mechanisms of these two phenomena—plastic deformation and parting—are unrelated. It will be shown below that plastic extension may be cut short by parting if the deformation is greatly accelerated, but there is no transition from the deformation to the failure. After the manner of its formation, a fracture is called a *brittle failure*, or, in better geologic terminology, a *brittle joint*.

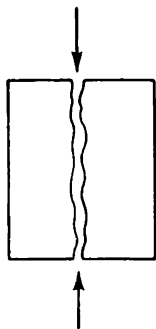


FIG. 233. Parting in compression.

Parting occurs not only in tension but also in compression. In the latter case, the fractures are parallel to the axis of compression and normal to the axis of lateral expansion (Fig. 233). Observations have shown that these compression fractures are formed when some critical magnitude of relative lateral expansion, which differs for different materials and conditions, has been reached.

Failure by shearing, which is produced by tangential stresses, as opposed to parting, is ordinarily directly associated with plastic deformation and is its continuation and ultimate outcome. The general features of the transition from plastic deformation to shearing, which has best been studied in the case of tension in metals, are the following: After elastic deformation, when the stresses have reached the threshold of plasticity, plastic deformation begins; this consists actually in a slippage of thin plates along the direction of the greatest shearing stresses. After a certain amount of strain hardening, as indicated above, the material is weakened; this weakening is associated with the formation of a neck in the body under tension (Fig. 232c). From the instant the neck is formed, a process of concentration of the plastic deformation in this narrower and narrower zone begins and develops rapidly. Study of metal specimens under the microscope (for example, Hanson, 1939) has revealed that at this time the plastic slip planes are at greater distances from each other. As the slip planes—which are the actual focus of the movement—become fewer and as they come together in a progressively narrower zone because of the concentration of the deformation, and also because in the zone of deformation the number of slip planes decreases, the velocity of the movement along each of these

planes, if the general rate of deformation is to be maintained, must inevitably increase. Ultimately there is shear along the surfaces on which the greatest movement was concentrated. Hence shear may be regarded as the result of a *concentration of plastic flow*, which was earlier disseminated over a great number of surfaces in a small number of planes or even one plane, leading to an acceleration of the slip and, finally, to a rupture of the body's continuity (Fig. 232d).

There is an interesting manifestation of the concentration of slip when the material undergoes creep. The creep curve showing elongation as a function of time in its full development, when the load is great enough, has four different parts. The last or fourth part, corresponding to acceleration of the deformation ending in failure, is the stage in which the slip is concentrated to the point of shearing, since it is characterized by the formation of a neck.

As the results of molding of shear deformations in highly plastic materials have shown (Beloussov and Kuznetsova, 1949), shear begins

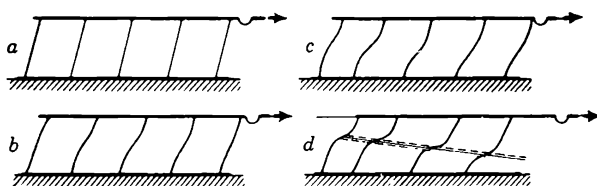


FIG. 234. The development of shear in a plastic material (petrolatum): (a) beginning of deformation—body is deformed uniformly; (b) beginning of slip; (c) development of this process; (d) shearing.

as a uniform deformation disseminated throughout the body. But at a certain stage the deformation becomes concentrated in the middle, and the other parts of the body are not deformed. This concentration ultimately ends in shear (Fig. 234). Theoretically, the shear fractures should appear at an angle of 45° to the axis of compression and extension, since these fractures are produced by the maximal tangential stresses. Normally, in actual solid bodies, however, including rocks, the angle between the shear fractures and the axis of the principal stresses is not 45° , and the angle formed by the shear fractures and the axis of maximal compression is less than 45° (from 30° to 45°). This deviation from the theoretical position is due to the composition and internal structure of the body and is evidently caused by friction along the fracture surfaces, since in shearing there are tangential stresses that produce a small amount of sliding of the walls of the fracture relative to each other. The actual position of the shear fractures is determined by the most suitable combination of magnitudes of the tangential stresses, on the one hand, and the friction along the fracture surfaces, on the other. Below, however, wherever such simplification is possible, it will be

assumed that the shear fractures follow the surfaces of the maximal tangential stresses.

Both in nature and under experimental conditions, the conjugate systems of shear fractures are not uniformly developed: one system predominates over the other. Often one system develops in the complete absence of the other. The reasons for this unevenness are apparently the lack of uniformity of both the material and the deformation; such heterogeneity is inevitable in every actual case. Because of this non-uniformity, the fractures of one system begin to appear earlier and thereafter become the locations of a progressively greater concentration of slip.

Geologic structures are excellent illustrations of the fact that, under the actual conditions in the earth's crust, the shear fractures are formed by a concentration of plastic slip. This mechanism of the formation of fractures is *plastic*: one may think of *plastic fractures* as opposed to *brittle fractures*. Shearing, however, may also be brittle when it directly follows elastic deformation without an intervening stage of plastic flow.

Thus there are two types of failure. Parting is associated with normal tensile stresses or lateral expansion in compression and is only brittle. Shearing is produced by tangential stresses and may be either brittle or plastic. In the latter case, it is the extension and termination of the process of plastic deformation and has the same mechanism.

Figure 235 shows the positions of the parting and shear fractures in small bodies undergoing various deformations. A knowledge of these is necessary for the interpretation of tectonic fractures under geologic conditions.

The two types of failure rarely occur in combination. Usually, under given circumstances, one observes only one type of failure, but with changing conditions it may be replaced by the other. This is determined not only by the material but also by all the external conditions on which the properties of the material depend: the temperature, the ambient pressure, the medium, the rate of deformation, and the stresses. These factors determine the type of failure and also its effect on the ultimate strength under different conditions. In the vast majority of materials, the ultimate strength in parting and shear is different. Depending on the conditions of the deformation, however, these limits increase and decrease differently in the scale of the stresses: under some conditions, the shearing strength of a given material is lower than the parting strength; under other conditions, it is higher. This complex relationship between the ultimate strengths explains the different behavior of bodies in failure; in each particular case, that type of failure will occur which requires the smallest stress.

Let us consider the effect of the conditions under which failure occurs on the failure itself. This discussion will be based on data obtained from a study of the mechanical properties of metals.

A rise in temperature lowers the shearing strength but has little effect on the parting strength. For this reason, when the temperature is raised, there is increased probability that the failure will take the form of shear. On the other hand, as the temperature is lowered, although the shearing strength increases, the parting strength shows almost no change; under these conditions failure by parting is more likely. By heating or cooling the same body, one may force it to fail by plastic shear or by brittle parting (Fridman, 1946; Uzhik, 1950).

External loads		Stresses		Type of fracture in rupturing	
		$+\sigma_{\max}$	$\tau_{\max}$	$+\sigma_{\max}$	$\tau_{\max}$
Tension					
Compression					
Shear					
Torsion					
Bending					

FIG. 235. Disposition of parting and shear fractures under various deformations (after Fridman, simplified in the case of shear).

The rate of deformation has an enormous effect on the material's properties. An increase in the rate has the same effect as a decrease in the temperature: the shearing strength increased and the parting strength remains almost unchanged. As a result, the material becomes more brittle and more likely to fail by parting than by plastic shearing. On the other hand, in a slow deformation the shearing strength decreases and plasticity may appear in the material. Thus at different rates of deformation, the same material may be brittle or may behave plastically.

Depending on the magnitudes and directions of the stresses, the relationship between the greatest normal and tangential stresses may

differ. Since parting is associated with normal and shearing with tangential stresses, the type of failure will also depend on the stresses in the body.

G. V. Uzhik (1950) has set forth the following criteria of the conditions for parting and shear. If τ_s is the shear strength and R is the parting strength, the condition for parting will be

$$\left. \begin{array}{l} \tau_{\max} < \tau_s \\ \sigma_1 = R_\sigma \end{array} \right\} \quad \text{or} \quad \frac{\tau_{\max}}{\sigma_1} < \frac{\tau_s}{R_\sigma}$$

and the condition for shear will be

$$\left. \begin{array}{l} \tau_{\max} = \tau_s \\ \sigma_1 < R_\sigma \end{array} \right\} \quad \text{or} \quad \frac{\tau_{\max}}{\sigma_1} > \frac{\tau_s}{R_\sigma}$$

The expression τ_s/R reflects the mechanical properties of the body under the given conditions of deformation. The expression $\tau_{\max}/\sigma_1$ is a function of the stresses in the body.

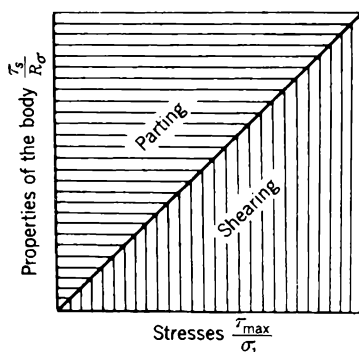


FIG. 236. Diagram of the type of failure as a function of the stresses in the body and its properties (after G. V. Uzhik).

Figure 236 is a diagram of the effect of the body's properties and stresses on the nature of the failure, which depends on both factors.

From the preceding formulas, we know that $\tau_{\max} = (\sigma_1 - \sigma_3)/2$. From this, it follows that if the difference between the principal stresses constantly increases while the absolute magnitude of the greatest principal stress shows little increase, shear will take place. If the maximal stress increases, the failure may be either parting or shear.

As shown above, when $\tau_{\max}/\sigma_1$ is sufficiently great, the failure will take the form of shearing; when it is sufficiently small, it will be parting. The magnitude of $\tau_{\max}/\sigma_1$, and thus the type of failure, depend only on the stresses in the body, since the magnitude of τ_s/R_σ is determined only by the properties of the body. Thus, for example, in uniaxial extension, $\tau_{\max} = \sigma_1/2$, or $\tau_{\max}/\sigma_1 = 0.5$. In biaxial and triaxial extension, $\tau_{\max} = (\sigma_1 - \sigma_3)/2$, or $\tau_{\max}/\sigma_1 = 0.5 - \sigma_3/2\sigma_1$.

Hence in biaxial and triaxial extension the probability of fracture increases, in comparison with uniaxial extension, since $\tau_{\max}/\sigma_1$ becomes smaller and the body behaves like a more brittle substance. On the other hand, in the case of a load with different signs or of biaxial and triaxial compression, because of the change in the sign of the second term in the right half of the last formula, the ratio $\tau_{\max}/\sigma_1$ becomes greater. This means an increase in both the plasticity of the material and the probability of shearing.

In torsion, when $\tau_{\max}/\sigma_1 = 1.0$, in most cases one may expect shearing.

The development of fractures, their direction, and the succession of phenomena depend to a great degree on whether the body contains any zonality, surface scratches, or any other defects or irregularities. There is a strong concentration of stresses about zones and scratches, and as the stresses increase, the zones also have a tendency to increase. Since the greatest concentration of stresses is around the sharp ends of fractures, this concentration increases in a straight line and extends the sharp ends of the fractures into the material to the extent that this is not prevented by the mechanical heterogeneity of the material or the distribution of stresses in the adjacent parts.

The mechanical properties of rocks that determine their deformation and failure under geologic conditions have been very little studied. There exist only a few fragmentary data on this subject (Birch et al., 1949; Hanson, 1939). In particular, it is known that the threshold of plasticity in marble and limestone increases with the increase in the ambient pressure and that the ultimate strength also increases under these conditions. For example, the ultimate shearing strength in compression at atmospheric pressure in the case of fine crystalline limestone is 2,600 kg/cm²; under an ambient pressure of 10,000 kg/cm², it reaches 13,000 kg/cm². The strength of quartz in compression under normal conditions is 25,000 kg/cm²; under an ambient pressure of 25,000 kg/cm², it reaches 150,000 kg/cm². For calcite, at an ambient pressure of 10,000 kg/cm², the ultimate shearing strength is twenty times greater than at 1 atm.

These data are completely inadequate. The entire problem as applied to rocks should be studied experimentally from the beginning. This problem is complicated by the great effect of the structure and texture of the rocks on their mechanical properties. It is known that rocks vary greatly in regard to structure, and this variation must be reflected in the variety of the mechanical properties. Since the structure of the rock is the result of the conditions of its formation and its earlier deformations, a study of the relationship between the properties of rocks and their previous history should be of great interest.

CHAPTER 26

THE MECHANICS OF TECTONIC FRACTURING

FRACTURES FORMED IN TENSION AND SHEAR UNDER NATURAL CONDITIONS

Tectonic fractures include both tensile joints and shear joints. We may first note the criteria for distinguishing these two types of tectonic fractures. If the rocks contain fractures and the displacement of the individual blocks consists only of movement of these blocks away from each other in the direction perpendicular to the fractures, the fractures are probably caused by normal tensile stresses. Rupture by shearing is produced by tangential stresses. Moreover, if the shearing is caused by tension or compression, the deformation is not normal to the fracture but oblique to it; if the shear fracture is produced by a couple, it will be formed in the direction of the shearing forces.

Shear fractures are the result and the termination of plastic deformation; consequently, rocks that have undergone disjunctive dislocations may frequently contain indications that the fracture was preceded by plastic deformation. In this case, the fracture results from a cohesive dislocation of the rocks. Sometimes, however, the plastic deformation is so small that brittle shearing takes place.

Most frequently the shear fracture is accompanied by displacement in the plane of the fracture; that is, it takes the form of a fault, which, in regard to the position of the fracture in space and the direction of the displacement along it, may, in geologic terminology, be defined as a normal fault, reverse fault, overthrust, strike-slip fault, or more complex displacement.

Figure 237 shows a layer of amphibolite enclosed within a gneiss and broken by transverse fractures into separated rectangular blocks. The fractures are in this case quite obviously tensile joints produced by normal stresses owing to the tension of the stratum.

Figure 238 shows another case of extension of a bed in which the latter is broken by oblique fractures at an angle of approximately 45° to the direction of the tensile stress. It is also important to note that the



FIG. 237. A layer of amphibolite enclosed within a gneiss and broken into blocks (*photograph by A. A. Sorskiy*).

bed underwent plastic contraction or necking before rupture, which was thus the final result of the plastic deformation. This is a typical case of rupture of a rock through shearing. The normal fault shown in Fig. 239, which is the ultimate development of a plastic bending of the beds, or a so-called flexure, is a similar case of a rupture.

Geologic literature, in addition to speaking of tensile joints, also mentions rupture. The present writer prefers the terminology used in mechanics, which avoids giving a double meaning to the word "fracture" and uses it only in its general sense to mean any kind of disjunctive dislocation.

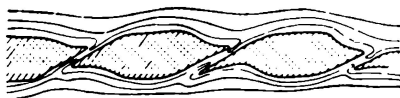


FIG. 238. A bed broken into blocks by shear.

Below we shall consider the conditions under which tensile joints and

shearing joints are formed, as well as those under which they are formed together. This discussion will classify the phenomena according to the mechanical action to which the rocks are subjected. We shall first distinguish fractures formed in extension, then fractures associated with compression, and finally fractures produced by shear. Fractures arising

in the course of bending will not be discussed separately, since the deformation of bending may be described approximately as unequal tension and compression in different parts of the body.

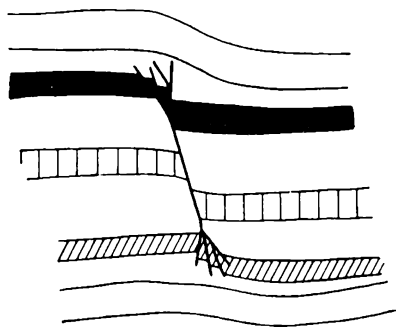


FIG. 239. Transition between normal fault and flexure, as seen in cross section.

beds is subjected to compression normal to the bedding planes, with the forces distributed as shown by the arrows in the diagram. The stresses exceed the plastic limit of beds *a* and *c*, which, in deforming plastically, spread out and decrease in thickness and by this means increase their horizontal dimensions. The material in them flows in the direction of the dotted arrows. In regard to the middle bed, the plastic limit of which is higher, there are two possible cases. In the first

TENSION FRACTURES

Let us first consider the conditions under which rocks may be subjected to tension. Figure 240 shows a group of beds in which a less plastic bed *b* occurs between two more plastic beds *a* and *c*. The entire group of

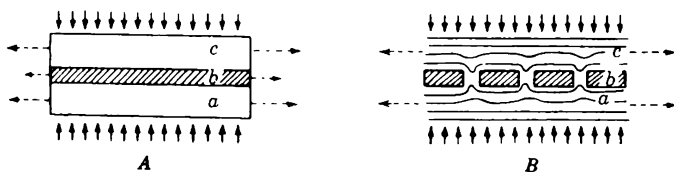


FIG. 240. Tension and fracturing in a less plastic bed between more plastic beds, under the conditions of normal pressure: (A) original state; (B) after deformation.

case, the stresses do not reach the plastic limit, and the bed is subjected only to elastic deformation. In the second case, the stresses exceed its plastic limit, but the material in this bed flows more slowly than in beds *a* and *c*, which are more plastic. In either case, beds *a* and *c* flow farther than bed *b* and extend beyond it in the course of their deformation. In the surface of the middle bed and tangential to it, there are tensile forces of friction. Bed *b* thus undergoes extension.

This case of extension of the rocks is very frequently encountered in

intensively deformed series composed of different formations. Many examples of such extension have been studied by A. A. Sorskiy (1950a,b) in the metamorphic series of Karelia, where the more plastic rocks are usually gneisses and the less plastic rocks undergoing extension are most often beds of amphibolites. Some specific examples will be given below.

The different variants of this case are shown by the diagram in Fig. 241. In the formation of continuous folds that are geometrically similar, the material in the beds is squeezed out of the flanks into the crests and troughs. This displacement of the material may differ in extent, depending upon the plasticity of the rocks; it is very intensive in the case of shales and is manifested to less degree in limestones or sandstones.

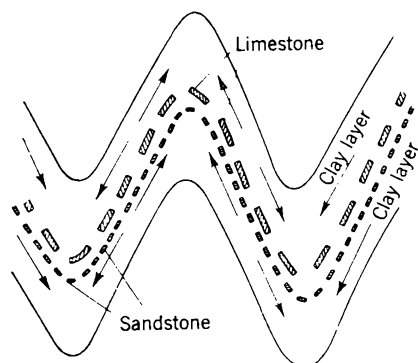


FIG. 241. Extension and fracturing of less plastic beds in a fold formed by more plastic rocks (diagram).

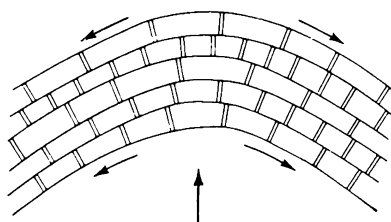


FIG. 242. Diagram of the tension in a bed within a domical uplift.

Observations in the field have shown that if a fold is composed mainly of plastic shales in which there is a considerable displacement of the material from the flanks into the crests and troughs, any interbeds of harder rocks (such as sandstones and limestones) will be stretched and broken into pieces (Kirillova, 1949).

Another example of extension in rocks is shown in Fig. 242. A group of beds subjected to tectonic pressure acting upon a certain surface in the vertical direction is bent into a fold. Since the beds at the margins of the zone of deformation remain in their previous locations, the surface of the beds that are bent becomes greater than it was in their original horizontal position. Thus the beds undergo tension. The bending and, consequently, the extension of any individual bed is due to pressure from the underlying strata, applied over its entire surface within the zone of deformation. This bed, in turn, transmits the pressure to the overlying bed and, in uplifting it, must overcome both the weight and the resistance to deformation of all the higher beds. Thus it is in a

state of compression between the overlying and the underlying strata, in the direction normal to this surface. It is clear that this case is in principle exactly like the preceding one.

The tension in all these cases is the result of forces applied over a fairly considerable area to the surface of the extended bed (distributed forces). Under tectonic conditions, no other case is possible: one cannot imagine conditions under which a bed would be isolated from the other beds and undergo extension resulting from forces applied at particular points or in particular sections. This feature of the mechanism of extension of rocks to a considerable extent determines the shape of the structures arising through extension.

The extension of a bed may have different results: there may be uniform plastic extension with a decrease in the bed's thickness, nonuniform plastic extension with the formation of necks in the bed (dividing it

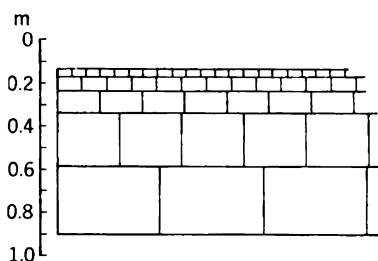


FIG. 243. Diagram of the relationship between the distance between joints and the thickness of the bed.

into separate pieces both by plastic and by brittle shearing fractures), and, finally, the bed may be divided by tensile joints.

To understand these different cases, let us consider the mechanism of jointing in a bed undergoing tension. Figure 237 shows a bed divided into separate pieces by tensile joints. In the process of extension, the pieces have moved some distance away from each other. This is the dislocation that was called *separation* earlier in this book. It is

notable that all these pieces are almost the same size. This is not due to chance: all other conditions being equal, the tensile stresses will break a thicker bed into larger pieces with corresponding greater distances between the joints, whereas a thinner bed is broken by more numerous fractures into smaller blocks (Fig. 243). The size of the blocks produced by tension in hard beds is a function not only of their thickness but also of the mechanical properties of the material. Beds of limestone and sandstone of equal thickness occurring between shales will be broken into pieces of different size. Sandstone, which has less strength than limestone, will form smaller blocks.

This rule associating the thickness and mechanical properties of the rocks with the dimension of the pieces into which the bed is broken when it undergoes extension between more plastic rocks suggests a comparison with the general distribution of jointing in rocks. It has already been shown that in stratified rocks the majority of intraformational fractures are normal to the bed, so that the jointing in general resembles brickwork. But the dimensions of the "bricks" are different in

different beds: they again depend on the nature of the rock, inasmuch as this affects the bed's mechanical properties, and on the thickness of the bed. Table 16 shows the results of a measurement of the average frequency of joints of this type in beds composed of one rock but of different thicknesses, as illustrated by the diagram in Fig. 243.

Table 16. Frequency of Joints in Sandstone Beds as a Function of Bed Thickness

I. Paleogene sandstones near Sochi (after Kirillova, 1949)						
Thickness, m.	0.03	0.05	0.1	0.25	0.3	0.4
Distance between joints, m.	0.04-0.1	0.05-0.2	0.1-0.3	0.2-0.4	0.4-0.8	0.4-0.6

II. Middle Carboniferous sandstones at Karaygyr in Bashkiria (after A. A. Bogdanov, 1947b)		
Thickness, m.	0.3-0.5	0.6-0.8
Distance between joints, m.	0.15-0.30	0.35-0.50

A work by A. S. Novikova (1951) on the jointing in the rocks of the Russian platform contains a diagram of the relationship between the frequency of the jointing and the thickness of the beds of carbonaceous limestones in certain parts of the platform. These data are shown in Table 17.

Table 17. Jointing in Carboniferous Limestones in Certain Parts of the Russian Platform

Thickness of bed, m.	0.15	0.4	0.5	0.6	0.8	1.0	1.2	2.0
Most common distance between joints, m. .	0.2	0.45	0.6	0.95	1.3	1.4	1.5	2.3

It can be seen that in every case the frequency of the joints increases and the distance between them decreases with the decrease in the thickness of the bed. The effect of the composition of the rocks is such that, when all other conditions are equal (especially in regard to thickness), the fractures are more frequent in softer rocks than in harder rocks. For example, a bed of argillite is broken by very numerous joints into thin plates perpendicular to the bed, whereas a layer of hard sandstone or limestone of the same thickness and in the same occurrence is divided into large blocks.

In general, there is no displacement of the blocks parallel to the plane of the joints. The displacement consists of a very small separation of the blocks from each other, opening the joints somewhat. Thus it is concluded that the general jointing is formed under the conditions of

extension in the bed but that the joints perpendicular to the bed are tensile joints identical to those produced by extension of a hard bed between softer ones.

Thus the relationship between the frequency of the joints and the mechanical properties of the bed and its thickness is a general rule applying to all cases of extension, and it is of fundamental importance in interpreting the phenomena associated with jointing. Another general rule is that, as a result of tension, the bed is divided not into two parts by a single joint, as would happen if the tensile forces were applied to any two points in the bed at some distance from each other, but into a large number of pieces separated by numerous parallel joints.

The mechanism of this phenomenon is most easily understood from the case in which the extension of the bed is caused by friction developing in its plane, due to the action of the more plastic beds on either side (see Fig. 240). The plastic beds are subjected to tension in such a manner that any two points on the surface of the bed move apart in the course of the deformation. This same process is transmitted by the friction to the intervening more brittle bed. Here, too, the tensile stresses develop along the entire surface of the bed, and any two points on this surface, whatever the distance between them, will tend to move apart. In other words, any point on the surface of the bed is a center of extension. It is important to stress, moreover, that any point is an *active* center of extension, inasmuch as the tensile stresses are applied to the entire surface of the bed. What is meant by "active" extension at any point will be easily seen from the following example.

Let us imagine that a bed is under tension from forces applied to its ends, whereas no forces are applied to the remainder of the bed. In this case, as long as the bed does not rupture but remains in a state of elastic tension, the elastic tension will be observed throughout the length of the bed and any two points in the surface will move apart. But as soon as the bed is ruptured, the picture is sharply changed. The formation of joints removes the tensile stresses in the bed, and if the forces continue to be applied, the two pieces of the bed that are separated from each other will gradually move in opposite directions. No other joints can form in the bed, since a single joint completely eliminates the stresses.

This situation is different when the tension is applied to the entire surface of the bed. Then a single joint, wherever it may appear, will not completely eliminate the tensile stress in the bed. These stresses remain even in the two broken pieces of the bed. For this reason, if the stresses are of sufficient magnitude, new joints will be formed.

How long can this division of the bed into pieces continue, and what determines the ultimate dimensions of the pieces? In Fig. 244, beds 1 and 3 are more plastic than bed 2 and, when subjected to compression,

are elongated in both directions from the 0 axis. As a result, the elements bordered by the lines a, b, c, d, e , and f (or a_1, b_1, c_1, d_1, e_1 , and f_1) are transformed into the elements bordered by $a', b', c',$ and d' , or, respectively, a'_1, b'_1, c'_1 , and d'_1 . It is easy to see that the farther one moves from the 0 axis, the longer will be the path traveled by the points in the first and third beds during the deformation: points a and a_1 are displaced to a smaller distance than points b and b_1 , and these, in turn, to a smaller distance than points c and c_1 , and so forth, as indicated by the horizontal arrows. Thus the velocity with which the third and first beds slide relative to the second increases as one moves away from the 0 axis.

Friction from the first and third beds acts upon the second. The friction depends on the rate of movement and increases as the latter increases. Consequently the forces of friction acting on the second bed are not the same at different points but increase with the distance from the 0 axis. In addition, each part of the bed is subjected to tension, since

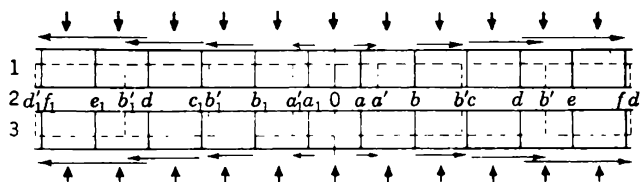


FIG. 244. Diagram illustrating the mechanism of fracturing of a bed in extension.

the end nearer to the 0 axis is acted upon by a smaller force than the farther end; in other words, the points farther from the axis in the second bed will tend to move more rapidly than those which are closer. At the same time, the smaller fragments of the bed are subjected to relatively smaller tension than the larger fragments, since the difference between the magnitudes of the forces acting on the ends will be greater in the larger fragment than in the smaller ones. This is the reason for the fact that, under any given conditions, there is a maximum critical size of the blocks into which the bed may be broken. This size is determined by the magnitude of the tensile forces and by the ultimate parting or shearing strength of the bed. For forces of a given magnitude, there will also be a critical limit to the size of the smallest pieces, below which the forces will be insufficient to create tensile stresses in the cross section of the bed greater than the ultimate strength of the rock. But in larger pieces of the same bed, under the same conditions, the tensile forces will be great enough that the stresses will exceed the rock's ultimate strength. For this reason, large pieces of the bed are likely to be broken up, whereas the smaller pieces are preserved.

These concepts contain the essence of the mechanism by which a

bed is broken into numerous small pieces in its extension; they also make clear the connection between the frequency of the joints and the composition of the rock and thickness of the bed. To produce rupture in a stronger rock, there must be correspondingly greater forces, which can arise only in a greater area of the bed. Hence the joints will be formed at relatively greater distances from each other. Similarly, the thicker the bed, the greater its cross section and the greater the tensile forces required to cause rupture. For joints to be formed under these conditions, the forces over a relatively larger part of the bed must be "collected"; this again increases the distance between the joints.

At the beginning of the process, a single joint will be formed whose position will be determined by the nonuniform composition of the rock. This first joint predetermines the positions of the others: their frequency is a function of the strength of the rock, the thickness of the bed, and the magnitude of the tensile stresses arising from the friction on the surfaces of the bed as it is stretched by the more plastic beds on either side of it. The lack of uniformity of the bed, which is inevitable at least in places, naturally affects the disposition of the joints. On the other hand, the fact that the average frequency of the joints is maintained throughout the deformation indicates that under ordinary conditions the nonuniformity of the lithology affects only the details of the process, which, in general, is determined by the rules described above.

Under natural conditions, what is the location of the axis of this deformation, which takes the form of outward flow of the plastic beds, and how far can this flow extend? In the dislocated regions of the earth's crust, areas in which the beds are compressed and areas into which the material squeezed out from them is injected are opposed to each other. In the first areas the beds are extended and stretched out, whereas in the second the material is accumulated in the form of disharmonic folds. Each area in which the beds are stretched out has a certain axis from which the material flows outward in both directions, although in the majority of cases it is extremely difficult to determine the position of this axis. The flow itself occurs within a very limited area, surrounded on all sides by areas in which the material is accumulated and injected between the beds.

The jointing pattern may also be used in the opposite manner: from the changes in the frequency of the jointing in a single bed, one may judge the changes in the magnitude of the forces that have acted upon the beds. The frequency increases in places where the deformation is relatively more intensive.

The Tectonophysics Laboratory of the Geophysical Institute, Academy of Science of the U.S.S.R., has conducted experiments that, on the one hand, have confirmed the general remarks made above on the mecha-

nism of joint formation in the extension of the beds and, on the other hand, have clarified a number of very important details of this mechanism.

I. V. Kirillova and Ye. I. Chertkova have used experimental laboratory models to study the process whereby a more rigid layer between two more plastic layers is broken into pieces. The samples prepared were parallelepipeds of gun grease with interlayerings of paraffin, representing a plastic rock series with interbeds of harder rock. The models were then compressed in the direction normal to the bedding surfaces. If the sample could be elongated along one axis, the gun grease was squeezed plastically along this axis whereas the paraffin layer was broken by transverse fractures into separate pieces. When the pressure was evenly distributed, the pieces into which the paraffin layer was broken were approximately the same. If the pressure was nonuniform, the pieces were smaller where the pressure was greater; this observation confirms what was said above regarding the mechanism of this fracturing. If the elongation took place equally in all directions, the paraffin layer was broken into many irregular angular fragments separated by lines of fracture running in various directions.

In another series of experiments, Chertkova (1950) made a study of fracture joints in paraffin and stearin. Circular disks of paraffin and stearin were bent upward, in imitation of tectonic uplifts (domes). The experimental conditions were as follows: a disk of paraffin or stearin about 2 cm thick was placed on a round base about 19 cm in diameter. The piston was moved upward by a screw mechanism and uplifted the central part of the disk, imparting a convex shape to it. In a number of experiments, the round piston was replaced by an oval or quadrangular piston, and the uplift acquired the corresponding shape. It appeared that the shape of the uplift affected the disposition of the fracture joints produced by the tension in the disk.

Radial joints are formed in a round uplift (Fig. 245). In the crest of an oval uplift, two longitudinal joints are formed parallel to the longitudinal axis of the uplift. These are connected by transverse joints, whereas the periphery of the uplift contains a number of radial joints. In the case of a rectangular uplift, the disposition of the joints is similar to the preceding case. Here there are also longitudinal joints parallel to the ridge of the uplift, whereas in the pericline there is a tendency toward the formation of radial joints. All these cases have their analogs in nature.

In all cases, the arrangement of the joints is determined by the direction of greatest curvature, and hence of greatest tension, since the fractures are perpendicular to this direction. If the curvature is approximately the same everywhere, as in a circular uplift, the arrangement of the joints will be symmetrical. The distances between them will then

be predetermined by the sum of the tangential tensile stresses and the breaking strength of the material.

In addition to the above joints, ring-shaped joints are often formed in the uplift and extension of the beds. It has been observed that tensile joints never intersect each other. They merely branch out; if there are joints in the same direction on the other side of the point at which the joints branch out, they never occur along the continuation of the previous joint, but are offset somewhat en echelon.



FIG. 245. Reproduction, in an experimental model of paraffin, of fracture joints in a round domical uplift (after Ye. I. Chertkova).

Important data were obtained by Ye. I. Chertkova on the development of the individual tensile joints. The joints on a round uplift, as a rule, are formed near the center and thence grow toward the periphery. For this reason, the outer ends of the joints are younger than their inner ends and are in an earlier stage of development. In addition, Fig. 245 shows one joint that is still far from complete (pointing toward the bottom of the figure). In tracing this joint from the younger end toward the center of the specimen, it can be seen that the first stage in the formation of a fracture joint is the appearance of very small embryonic fissures, which are short and unconnected but located within

a single zone and oriented in a single direction. They are en echelon with uneven edges and have a vermicular curvature (Fig. 246a). The next stage is the growth of embryonic fractures along the strike on both sides; this is undoubtedly due to the concentration of stresses around the sharp ends of the original fractures. As the embryonic fractures grow, they approach closer to each other, but because of their en echelon arrangement they usually do not come into contact, since their ends are usually parallel (Fig. 246b). The next stage is the most interesting: the fractures whose ends are next to each other appear to grow together. As it grows, the end of each fracture bends toward the next fracture, so that the two are joined as shown in Fig. 246c. This behavior is again explained by the concentration of stresses around the points of dislocation. Wherever the fields of concentrated stresses associated with adjoining fractures approach each other, the narrow strip between them is broken.

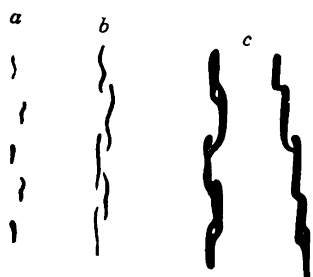


FIG. 246. Diagram showing the growth of a tensile joint: *a*, *b*, and *c* are stages of the process.

The large fissures formed in this manner from the combination of small embryonic fractures thus have certain morphologic peculiarities. In the first place, they can be



FIG. 247. Sketch of a system of natural tensile joints (after Ye. Ye. Zakharov).

divided into separate broad areas arranged more or less like steps and into narrower connecting strips at a certain angle (sometimes perpendicular) to the strike of the fracture. This can be seen in Fig. 245. In the second place, wherever the embryonic fractures have been joined together in the manner shown in Fig. 246c, the rock forms a narrow wedge or lens bordered on both sides by fractures. These peculiarities of the morphology of tensile joints are common in geology, and any field in which these joints are developed will provide numerous examples (Fig. 247). Thus it has been established that certain essential features of tensile joints are determined by the concentration of stresses near these zones, and especially near the sharp ends of already existing fractures. This interpretation of the formation of tensile joints is also fully applicable to the other cases of tension in the beds. Figure 248 shows a bed that has undergone not brittle rupture but plastic extension, with the formation of numerous necks repeated periodically. These are usually called *tectonic lenses*. The frequency of these lenses, like the frequency of tensile

joints, decreases, and the distance between the lenses increases from more plastic to more brittle rocks and with the increase in the thickness of the bed. For the formation of a neck to occur, the tensile stress must exceed the threshold of plastic deformation of the given bed as a whole. But this increases with the increase in the dimensions of that part of the bed which is subjected to tension. In this case, the frequency of necking is a function of the same factors as those which determine the frequency of the tensile joints, if the plastic limit is substituted for the ultimate breaking strength.

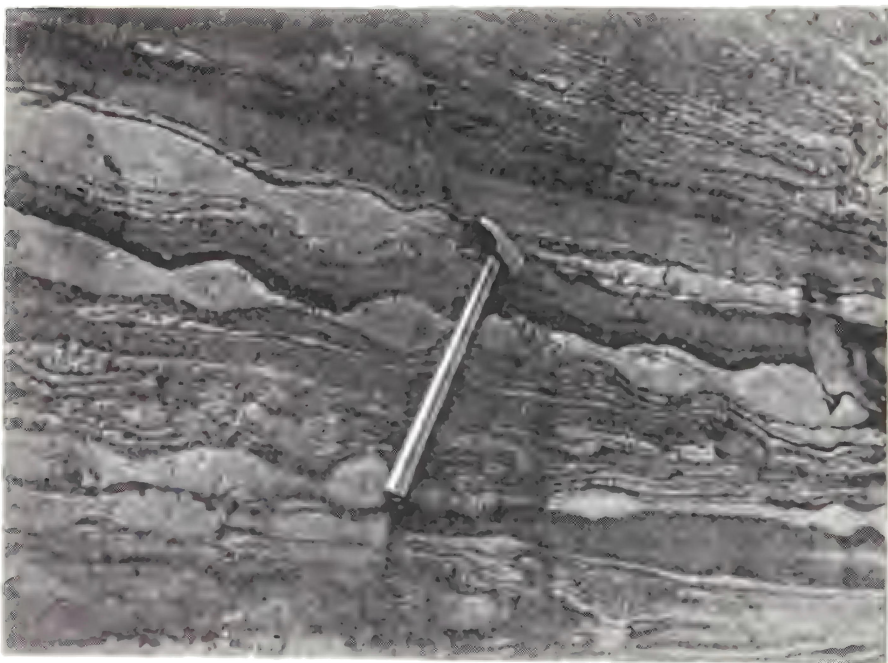


FIG. 248. Tectonic lenses in a pegmatite enclosed within gneisses (*photograph by A. A. Sorskiy*).

Certain beds are extended without the formation of necks, and they retain their uniform thickness. During the course of their deformation, such beds must be in a very plastic state; their plastic limit is so low that they can be deformed by very small tensile stresses between two nearby points on the bed. Consequently such beds, under the conditions of the deformation, have properties close to those of a viscous liquid.

If the plastic deformation goes through its full development, it ends in rupture resulting from the concentration of plastic slip and the acceleration of its rate. Once shear joints have replaced the necks, the bed is broken into pieces. But whereas these pieces have rectangular

shapes if the bed is broken by tensile joints, when it is broken by shear joints oblique to the bedding planes, the pieces have rhombohedral or (in the case of necking) lenticular shapes.

Beds broken into rhombohedra or lenses arranged en echelon are encountered very frequently. Figure 249 shows a block diagram drawn from field observations, showing that a bed of limestone between more plastic beds of marl has undergone extension and has been broken into pieces separated by shear joints in two directions: along the dip and along the strike. But the dimensions of the pieces in these two directions are very different: they are much larger in the section along the strike than along the dip. This indicates that the tension in the direction of the dip was several times greater than the tension along the strike: in order to fracture the bed in the direction of the dip, it was sufficient to "collect" the forces of tension from a relatively small surface area of the bed; to break the same bed into pieces along the strike, the forces had to be "collected" from a greater area. The ratio of the frequencies of the joints in the two sections indicates the quantitative relationship between the magnitudes of the tensile forces in these sections. If the frequency is greater and the distance between fractures very small, the phenomenon resembles cleavage in the case of both shear and tensile joints.¹

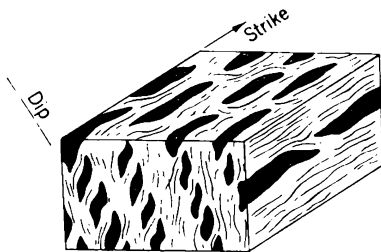


FIG. 249. Block diagram of outcrops in a suite of marls and limestones (black) of Devonian age in the Kara-Tau Range.

In all cases just described, the shear joints may be either plastic or brittle. Brittle joints are distinguished from plastic joints by the absence of any plastic deformation preceding the rupture.

Since the mechanism of the formation of tension fractures has been illustrated with examples of small-scale disjunctive dislocations, the reader may have gained the impression that tension is reflected only in such minor forms. This, however, is not the case. Tension over large parts of the earth's crust results in the formation of very large fractures that usually take the form of normal faults. The next chapter will present some examples of great faults associated with tension in the crust.

All the fractures described above are of tectonic origin, formed as the result of external tectonic forces applied to the bed of rock. Some authors call such fractures *exokinetic* (Ivanov, 1939). But rocks also contain fractures of nontectonic origin that are formed by a mechanism

¹ Some researchers, especially coal geologists, refer to all joints as cleavage (Ivanov, 1939). This is incorrect, since cleavage is a particular phenomenon intermediate between plastic deformation and fracture and should be considered as such.

very similar to that of the formation of tectonic fractures. For example, rocks may be fractured in the course of creep, in which they are subjected to tension. Such tension arises because the force of gravity draws some part of the mass of rock down a slope and because this movement is resisted by friction developing in the surface of slip during the creep. The friction is distributed throughout the surface and combines with the force of gravity to produce the tension. Thus both tensile joints and shear joints may be formed in a mass undergoing creep.

Rocks are also fractured through decreases in their volume. The volume of igneous rocks decreases as they solidify and the volatile constituents escape from the magma; sedimentary rocks decrease in volume through desiccation in various physicochemical processes (such as coal formation). Moreover, a decrease in volume always involves the appearance of tensile stresses in the rock; these may be the direct cause of the fractures that have been called *endokinetic* (Ivanov, 1939). The tensile stresses arise because the decrease in volume is resisted by external friction, which is thus a reaction and combines with the internal forces of compression in the rock to produce the tension.

Lava, for example, once it has been extruded upon the earth's surface, tends to contract. If no friction were developed between the lava flow and the earth's surface, the lava could decrease its volume in a continuous process, without disrupting its continuity. Because of the friction, this is impossible, since the particles of the lava flow that are at some distance from the center of contraction must do a great amount of work to overcome the friction. As a result, many separate centers or axes of contraction appear; only the directly adjoining particles move toward these centers as the volume decreases. Thus, as a basalt flow solidifies, it is broken up by columnar joints and the columns of basalt are separated by tension fractures. The cavities formed by these columnar joints correspond to the decrease in the volume of the basalt.

Since friction opposes the contraction of the basalt, the force of friction acts in such a manner that the basalt is subjected to tension. Hence there is no difference in principle between the formation of tensile joints or tension fractures through tension in a rock and the formation of the same fractures as the rock decreases its volume in cooling. The volumetric forces existing within the basalt are "collected" throughout the volume, and the dimensions of the basalt fragments are thus determined by the relationship between the magnitudes of these forces and the friction, on the one hand, and the parting strength of the rock, on the other.

Tension fractures resulting from decreases in the volumes of rocks are very common in nature. They have also been observed in intrusive rocks. It is known that the fractures in intrusives are regularly disposed with relation to the flow structure. As regards the conditions of forma-

tion of these fractures, the following may be said: the minerals were oriented during the stage of flow, while the igneous rock was still in a hot, plastic state, and still contain its volatile components. With further cooling and crystallization, the volume of the rock decreases, producing tensile stresses equal to the external tensile forces acting on the rock. The reaction to this tension is the formation of fractures. The disposition and frequency of the fractures in different directions are a function of the anisotropy produced by the *flow structures*. Hence the connection between the directions of the fractures and the principal axes of tectonic deformation is not a direct one: the immediate result of the strain is the creation of flow structures in the intrusive, and these in turn govern the anisotropy of the material. The latter is reflected in the formation of tension fractures with greater or less frequency in specific directions, depending on the strength of the rock in those directions.

Ordinarily the longitudinal fractures, parallel to the flow structures or the grain of the rock, occur with greater frequency and the distances between them are relatively smaller. Transverse fractures normal to the grain are located at greater average distances from each other. Hence the fragments formed by them are analogous to bricks, whose short sides are normal to the flow or grain and whose long sides are parallel to it. This distribution of the fractures agrees fully with the greater longitudinal strength of the rock in the direction of the grain, established from many examples, as compared with its strength in the lateral direction.

It is known that regular systems of joints are usually not observed in intrusives with no particular orientation of the minerals, in which the joints are predominantly curved. Homogeneous granites, for instance, often contain so-called ellipsoidal joints forming flattened ovals. These curved joints testify to the uniformity of the rock, which is almost quasi-isotropic. The contraction toward certain centers is more or less the same in all directions.

Fractures resulting from desiccation similarly arise in sedimentary rocks. When clay dries, its volume decreases and it is broken up by polygonal joints originating in exactly the same manner as the columnar jointing in basalt.

Thus the mechanisms of the formation of both tectonic and nontectonic joints (the latter resulting from creep or from a decrease in the rock's volume) have much in common. The difference is that the tectonic fractures arise from external forces. They are determined by the nature of the deformation and by the relationship between one bed and the next; the relative plasticities of adjacent beds, for example, play an important role in the formation of these fractures. Nontectonic fractures are formed mainly by the force of gravity, the contraction resulting from a decrease in volume, cooling or desiccation, and so forth. Pre-

ceding deformation plays a role in determining the anisotropy of the material and thus affects the directions and frequencies of the fractures.

One more fact must be noted. The majority of rocks, especially those with great strength (crystalline rocks, dense limestones and sandstones, extrusive rocks, and especially volcanic glass) occur below the earth's surface in a state of volumetric stress. One indication of this state is the behavior of the rock when specimens are broken off from it; the small pieces of rock chipped off by the geologist's hammer break away with an energy clearly much greater than that of the hammer blow itself. Every geologist knows the whistling sound these fragments make as they fly through the air. Other indications are found in the behavior of rocks in mining or quarrying operations, which sometimes produce sounds like explosions accompanying the sheeting of the rocks. The rock, recently freed of the surrounding mass pressing in upon it, expands suddenly and shoots out large, far-flying pieces from its fresh surfaces. These pieces cannot be returned to their former positions, since their volume has expanded as they were broken off (Slesarev, 1948).

Thus the rocks beneath the surface are in a state of stress resulting from the overburden of rocks above them. Since their compression is volumetric, it is elastic and the stresses may be retained for an indefinite time. They are liberated, however, as soon as there is a local removal of the load, which may be due to mining operations or to natural processes of denudation. This removal of the load by natural processes must be the reason for the formation of the joints, which may be observed everywhere, that are roughly parallel to the earth's surface. As the surficial parts of the rocks expand, they break away from the deeper parts and form such joints.

COMPRESSION FRACTURES

It has been shown above that compression of a brittle body sometimes results in the formation of fracture joints whose strike follows the direction of the compression. It is generally considered that, under geologic conditions, the formation of parting joints in the compression of rocks is very common. It may also be remarked that all cases in which the formation of parting joints is associated with compression, but not with tension, deserve special study and require particular confirmation. There is some basis for doubting the commonly accepted view of the widespread distribution of such cases. It is not impossible that in many cases in which the parting joints are considered to have originated from compression parallel to the fracture, they have actually been formed by tension normal to it. Such misunderstanding is all the more possible because of the erroneous conviction of many geologists that the compression of a rigid body in one direction and its extension in the

perpendicular direction are mechanically equivalent. In reality, the stress states that affect the mechanical properties of the body are different in these two cases. In linear compression, as long as the deformation remains elastic, the body decreases in volume. This difference is reflected in the body's properties in such a manner that the body has greater plasticity in compression than in tension.

As a rule, uniaxial and biaxial compression results in the formation of oblique shearing joints. These are either the ultimate termination of plastic deformation (in the case of plastic shear) or else the result of immediate brittle rupture. Thrusts and strike-slip faults are examples of tectonic shearing joints arising from compression. Under geologic conditions both are, in the overwhelming majority of cases, plastic and closely associated with the plastic deformation preceding them. The plastic deformation, in thrusts, is reflected either in folding or in vertical flexures, whereas a strike-slip fault is thus associated with a horizontal flexure—that is, with a step-wise bending of the beds in the horizontal plane.

It was indicated earlier that thrusts and strike-slip faults disappear in all directions. According to observations, the plane of a thrust or strike-slip fault in the area where it dies out spreads into many branches and is gradually replaced by scattered plastic deformation (Fig. 239). From this, it is easy to imagine the development of an overthrust or strike-slip fault as a plastic shear fracture. First, because of a certain concentration of the slip, a zone of intensified plastic flow or, more accurately, of higher gradients of the rates of such flow is formed. The structural reflection of this zone is a bending of the beds. A further concentration of slip leads to a stretching of the beds in this zone and to the appearance in it of very small embryonic fractures oriented parallel to the movement. Thereupon these fractures grow together and the number of "working" parallel fractures is finally reduced to one, along which all further tectonic movement takes place.

In nature one can observe only the result of this process, whereas in laboratory experiments one may trace its course. Figure 250 shows the results of an experiment in producing shearing joints grading into a thrust, under the conditions of compression. It can be seen that the lateral compression of gun grease has produced shear fractures that die out in both directions in the vertical section, merging into a system of branching fractures and grading beyond these into a clearly reflected

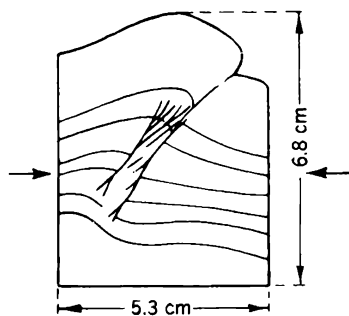


FIG. 250. Experimental reproduction of a thrust grading into plastic bending of the beds.

bending of the layers. Here we may see, in "frozen" form, the concentration of slip: the individual parts of the fractions are in different stages of formation; in the middle the process of concentration is already completed, whereas at the ends it is still developing. Thus one may observe a transition from plastic bending, through small fractures, to a single large fault. Under natural conditions as well, the ends of thrust or shear fractures are younger than the middle parts, so that in tracing such a fracture from the middle to its ends one passes through the various stages of its development.

Under the conditions of considerable symmetrical compression in these experiments, two slip planes were formed in different directions

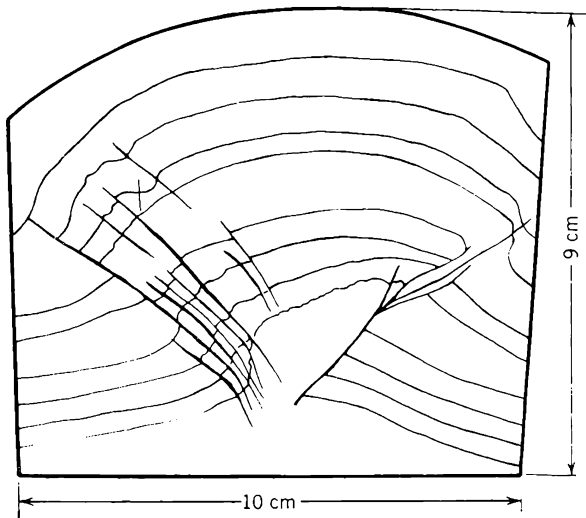


FIG. 251. Repeated shearing in a stratified plastic material under compression.

and the wedge-shaped block between them was squeezed upward. Figure 251 shows a section through a number of layers of petrolatum, also compressed horizontally and cut through by a series of shearing surfaces moving toward each other. Because of the layering, the structure of the specimen is clearly visible. From the undulations in the bedding planes one may see that the shear planes are not only those that can be seen by the naked eye. There is a considerable number of shear planes that are just beginning to form through concentrations of the plastic flow and are still apparent only in the sharp bending of the layers.

Movement along thrust fractures is the result of compression of the body in a particular direction and is an expression of this compression. Compression of the same magnitude may be obtained either by a considerable displacement along a single fracture or by smaller displace-

ment in each individual case along many smaller fractures. Experiments have shown that if the horizontal compression occurs under such conditions that the upward squeezing of the material is unhindered, a single fracture or a small number of fractures is usually formed and the displacement along each of them is relatively great. But if the upward pressure encounters resistance from the overlying material, because of weight and resistance to deformation, the latter spreads farther from the source of the pressure and numerous shear fractures are formed with relatively little displacement along each of them. In Fig. 252, along the interlayer of powder, one may see a large number of shear fractures with small displacements in the lower layer, which is deformed in a medium in which the upward movement is hampered by the overlying layer of harder material (paraffin).

In the compression of rocks, because of the fact that shear fractures arise as the result of concentration of the plastic slip, lensing frequently takes place as an intermediate stage in the deformation. This lensing fully resembles the similar phenomenon described earlier in discussing the deformations associated with tension. It is best observed where the beds are compressed normal to the bedding planes. The lenses are bounded by zones of intensified slip, corresponding to the future system of conjugate shear fractures.

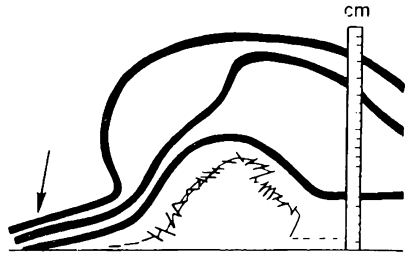


FIG. 252. Petrolatum (white) with interbeds of paraffin (black), pressed downward at the left (arrow) and squeezed horizontally at the right.

This case of the formation of lenses may be easily reproduced on models. Ye. I. Chertkova prepared a sample composed of three layers of petrolatum. The layers each had different colors and (from the bottom upward) thicknesses of 20, 15, and 15 mm, respectively. The length of the specimen was 110 mm, the width 55 mm, and the total thickness 50 mm. In the horizontal position the specimen was compressed vertically between two boards, as a result of which its thickness decreased to 27 mm and its length increased to 195 mm. The width remained as before, since the elongation of the specimen in this direction was prevented artificially. Figure 253a shows that the uppermost and lowermost layers of petrolatum during the course of their elongation underwent clear lensing, with the formation of swollen and squeezed-out portions. The middle layer was subjected to less degree of lensing, but it acquired a wavy surface. As the compression continued, one could first observe the appearance of indistinct zones of intensified flow bordering the sharp ends of the lenses and then the formation of distinct shear fractures rhythmically dividing the specimen into prisms (Fig. 253b).

Important data on the mechanism of development of shear fractures may be obtained by observing the changes taking place in the top surface of a prismatic specimen of plastic material in the process of its compression. By examining this surface through a magnifying glass under oblique light, one may observe a system of very fine lines formed by the mutual intersection of rhombs stretched out along the long axis of the specimen, perpendicular to the compression (Fig. 254). These lines are directly related to the mechanism of formation of the shear

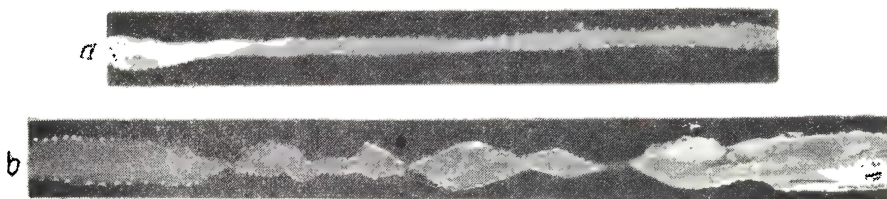


FIG. 253. Formation of lenses in a layer of petrolatum under compression; the axis of compression is vertical in the plane of the illustration.

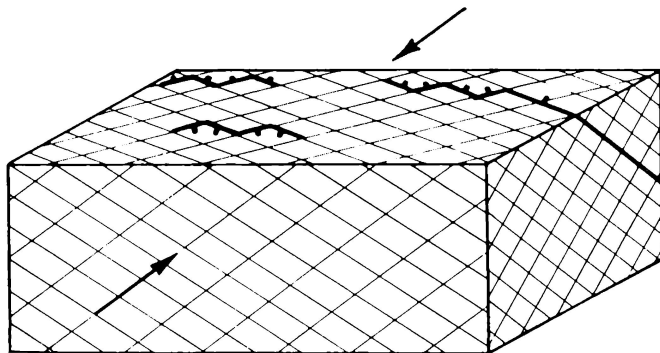


FIG. 254. Grading of Lüders lines into shearing joints in a plastic material (petrolatum) under compression.

fractures. The latter appear initially in the form of short surfaces of displacement extending in the horizontal projection generally perpendicular to the axis of compression, but with a toothed shape, since they are associated with individual sections of these intersecting lines. Thereafter, with a further development of the deformation, the slip surfaces grow in length. When the displacement along these reaches such an extent that a slip surface may be observed in the bottom of the hanging wall, this slip surface appears furrowed, corresponding to the toothed shape of the upper edge of the fracture. Hence it is clear that the intersecting lines mentioned above are signs of the emergence at the

surface of the tiny slip planes produced by the deformation of the specimen.

SHEAR FRACTURES

This section will consider some of the problems in the mechanism of formation of tectonic faults that are produced by the action of a couple on the rock and are thus the result of a shear deformation (in the mechanical sense). Such a couple arises, for example, in the vertical uplift of one part of the earth's crust relative to another or in a similar horizontal displacement. The following chapter will present examples of vertical relative uplift of blocks of the earth's crust along tectonic faults. This relative uplift may be due to the fact that tectonic forces of uplift of different magnitudes act on the individual blocks. Examples of horizontal tectonic shear dislocations will also be presented.

Shear results in the formation of both tensile joints and shearing joints. Their positions with relation to the directions in which the forces are acting may be determined from Fig. 235. Under shear, the rectangle is transformed into a rhombus (the deformation is considered as occurring in a single plane). The long axis of the latter indicates the direction of greatest tension, and the short axis the direction of greatest compression. The planes of maximal tangential stresses are parallel to the two sides of the rhombus. The directions of the principal axes and the planes of the greatest tangential stresses are determined by the positions of the tensile and the shearing joints: the tensile joints are normal to the axis of greatest tension; of the shearing joints, one system is parallel to the direction of the couple and the other perpendicular to it. These will be the positions of the shearing joints if it is considered that they are inclined at an angle of 45° to the principal axes.

Tensile joints associated with shear deformations are widespread in nature. Usually they are easily detected by their disposition en echelon (Fig. 43). Such echelons have been mentioned earlier: the right echelon is associated with right shear and the left echelon with left shear. In the case of small dislocations, the fractures may be very short; when the dislocations are greater, they may sometimes be several kilometers in length. They are typically lenticular in shape and thicker in the middle, where the shear displacement was greatest. S-shaped bending of the fractures has also been observed (Shainin, 1950). The latter occurs when the earlier formed part of the fracture is rotated during the process of continuing shear deformation, whereas the new parts of the fracture at its ends continue to grow in the directions determined by the same original directions of the principal tensile stresses.

In the development of tensile joints under shear conditions, the joints will have a given frequency, which may differ under different condi-

tions but which on the average is maintained throughout. This frequency is apparently determined by the same mechanism that determines the distance between the tensile joints under the conditions of tension (see above).

Shearing joints under shear conditions (in the mechanical sense) are most often formed parallel to the direction of the shear. In geology they take the form of upthrusts and normal faults when the movement is vertical and of strike-slip faults when the movement is horizontal.

The faults that occur in connection with shear deformation may be studied with the aid of models. In one series of experiments, tensile joints en echelon were obtained under shear (Fig. 255). Other experiments were carried out to obtain shearing joints under shear conditions.

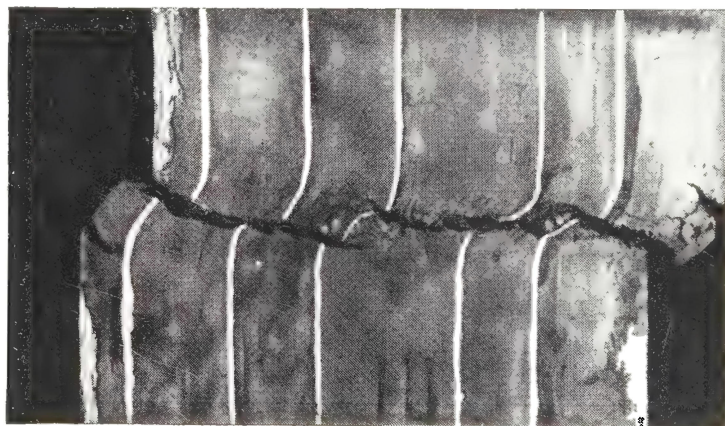


FIG. 255. Reproduction of tensile joints in the process of shear (*experiment by M. V. Gzovskiy*).

The latter were conducted with the same apparatus that was used earlier in modeling tensile joints: a cylindrical vessel with a circular piston in its bottom. A layered specimen of gun grease or petrolatum was placed over the bottom of the apparatus. The upward pressure from below in the central part of the specimen was produced not directly by the hard piston but by a plug of paraffin or petrolatum placed upon the piston. As a result, the middle part of the specimen located over the piston was pushed up and uplifted. During the process, the following peculiarities were observed in the course of its development: At the beginning of the deformation, the layers of the specimen were elevated over the piston, bending plastically and forming a dome. The diameter of the dome, as we already know, always increases upward from below, so that if, in a section through the center of the dome, lines are drawn connecting the points at which the layers bend from their horizontal to their inclined positions—or the foot of the dome in the various

layers—these lines will diverge upward and usually form an angle of 60° from the horizontal (Fig. 256).

In connection with the upward expansion of the dome, the bending of the lower beds in the flanks is sharper than the bending in the upper layers. Moreover, whereas the lower layers in the portion immediately above the piston retain their horizontal attitude, forming a box-type uplift, the upper layers show a transition to a very gentle domical bending. Because of the steeper bending in the lower layers and their greater thickness over the piston, the zone of bending is squeezed and the bedding planes are much closer together. This stage of the deformation may also be seen in Fig. 256. These zones of squeezing are

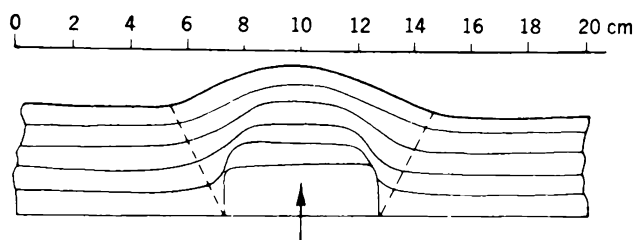


FIG. 256. Domical bending of layers of petrolatum uplifted from below by a plastic die.

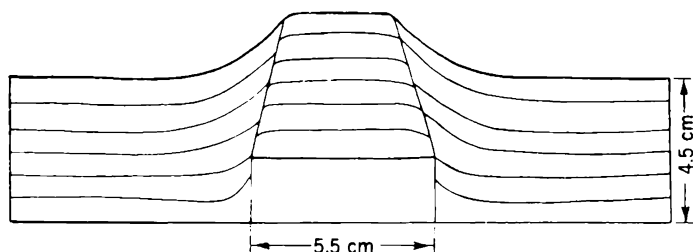


FIG. 257. Continuation of experiment shown in Fig. 256.

not vertical but inclined: they converge upward, usually forming an angle of 70° to 75° with the horizontal plane. The zones in which the thickness of the layer is diminished are also zones of intensified plastic slip; thus as the piston is raised still further, shear fractures form along these very zones. These always grow from the bottom up and gradually penetrate the higher layers.

If the specimen has a certain thickness relative to the diameter of the piston, the upward-growing fractures will emerge at the surface (usually at different times in different sides of the specimen). In cross section (Fig. 257) it can easily be seen that the amplitude of the displacement along the faults decreases upward from below. On the basis of the supposition that shearing faults are the result of a concentration of slip,

this change in the amplitude of the displacement along the fault is due to the fact that, with increasing distance from the surface at which the pressure is applied, the equivalent plastic slip is disseminated over a greater area. This dissemination of the slip as one moves upward may also be observed directly in sections through the model. Figure 258 shows that the fault on the left branches upward. This is identically the same phenomenon as the dissemination of shear fractures produced by compression.

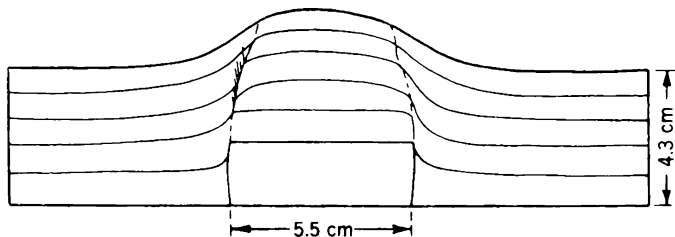


FIG. 258. Branching and dying out of shear joints.

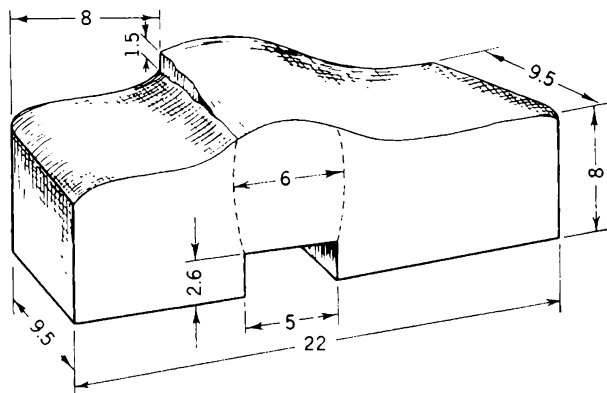


FIG. 259. Barrel-shaped cross section of a part of a plastic bed compressed above a die; material is petrolatum.

In discovering the laws governing the formation of faults, it is important to establish the disposition, under various conditions of deformation, of the shear joints arising as a result of the concentration of slip: will they always be arranged as in the experiments described immediately above, converging upward, or can they also be disposed otherwise? Experiments have shown that these planes converge upward, forming an angle of 70° to 75° with the horizontal, as already mentioned, in specimens of small thickness (to 4 to 5 cm when the diameter of the piston is 5 cm). But in specimens of greater thickness (more than 7 to 9 cm with a piston diameter of 5 cm), the faults produce a different shape:

a barrel shape is obtained, bordered by faults that first diverge upward in the section and then begin to converge again (Fig. 259). These faults remain very steep (about 70° to 75°).

Since the shapes of these shear faults must repeat the shapes of the

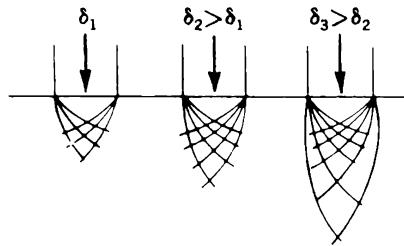


FIG. 260. Slip lines in downward pressure of a die in a plastic medium.

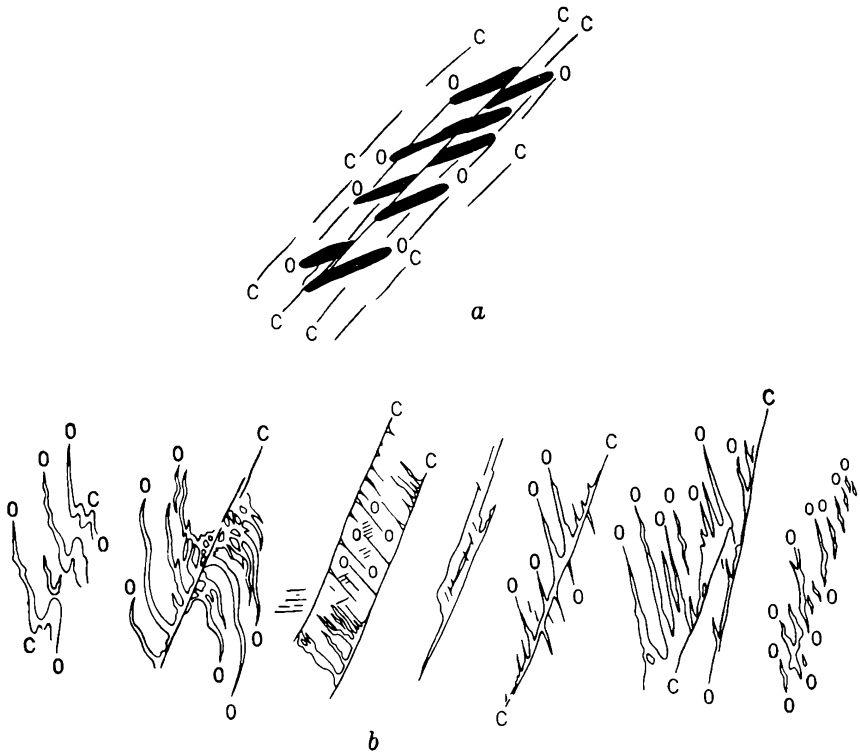


FIG. 261. Combination of tensile joints (o) and shearing joints (c) in shear.

planes of plastic slip, it is interesting to compare them with the most appropriate system of such planes. Figure 260 shows such a system, produced by the impression of a die into a plastic plug (Gubkin, 1947). The sketch shows the individual stages corresponding to the successive

development of the process. At the beginning, when the deformation has still not penetrated deeply, the slip planes proceeding directly from the die converge. In the later stages they first diverge and then converge again, forming a barrel-shaped figure.

In all these cases, the shearing fractures develop following a more or less considerable plastic deformation. But when, under similar conditions of shear, the shearing joints appear by way of brittle dislocation, their shapes are governed by the same rules.

In conclusion, we may dwell momentarily on the combination of tensile joints and shear joints produced by the same shear deformation. Figure 261*a* shows a structure that includes joints of both types. They may be combined in any succession: in some cases, the shear joints may precede the formation of the tensile joints; in others, the tensile joints may be formed earlier than the shear joints. The first will be the usual case in slow deformation of a plastic material, when the ultimate shear strength is less than the ultimate breaking strength. The second case will be more characteristic of brittle rocks and more rapid deformation, when the resistance to plastic deformation increases and the ultimate shearing strength is greater than the parting strength, which is known to have little relationship to the conditions of deformation.

Of special interest is the case in which tensile joints are formed first and then shear joints. In that case, in the zone of the future upthrust or strike-slip fault, a series of short en echelon tensile joints is formed, and these open wider with time. As the deformation develops, these joints gradually begin to be connected to each other by very small shear joints, which finally merge into a single fracture or zone of fractures.

CHAPTER 27

TECTONIC CONDITIONS PRODUCING FAULTS

The preceding chapter has shown that tectonic ruptures are produced by various mechanical forces: tensile, compressive, and shear. Each of these forces may result in the formation of both shear fractures and tension fractures. In all cases, the tension and compression are understood as being parallel to the beds or in the horizontal direction.

This chapter will discuss the tectonic conditions under which individual parts of the earth's crust undergo tension, compression, and shear and which produce the most typical combinations of fractures and determine their position in the general course of development of the earth's structure.

TENSION

Tension and Bending Fractures

The preceding chapter reviewed fracturing caused by tension. The fractures were formed by the dragging and stretching of rigid rocks enclosed in plastic rocks on tectonic uplifts. In addition, there must be some mention of step veins resulting from tension. Such veins have filled the fractures formed in hard rocks that were stretched when the surrounding relatively plastic beds were deformed. This type of fracture occurs, for example, in hard dikes enclosed in shales, sandstones, and other more plastic rocks if they are compressed normal to the bedding planes. Figure 262 shows dikes cut by step veins. Not all the veins pass through the dikes; many extend only partially into them.

Here we shall review in more detail the fracturing related to the formation of domes, arches, uplifts, etc. These structural forms inevitably cause tension in the earth's crust, since they are directly formed by vertical forces. This point is

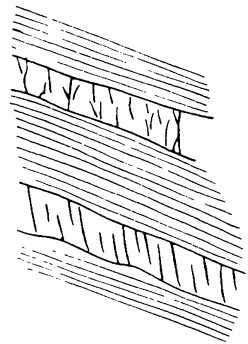


FIG. 262. Step veins in dikes enclosed in shales.

very important: if local vertical forces elevate a part of the earth's crust, that part is subjected to stretching and the area of the upwarped beds is increased. The stretching may produce fractures and cause broad and deep cracks in the earth. In structural uplifts such as continuous folds, produced by horizontal compression, no tension fractures apparently occur. Stretching may produce both fractures and faults. First we may review the fractures formed in the crests of uplifts, which are therefore called *bending fractures*. Both shear and tension fractures may be formed by bending.

It must be said that stretching occurs not only by upwarping of the earth's crust but also by downwarping, if both movements are caused by vertical forces. Fractures resulting from subsidence, however, do not occur as widely as those resulting from upwarping: subsidence and deposition of younger sediments bring the earth's crust into the zone of higher pressures, which prevent the formation of fractures. Upwarping caused by vertical forces which, in turn, causes the earth's crust to be stretched may vary widely in intensity. There may be large and small discontinuous folds, gentle and broad antecises, or even anticlinoria in geosynclines, provided the upwarping forces in the earth's crust are sufficient to produce the stretching. Stretching produces both shearing and tension fractures. The fractures frequently develop into faults, which will be discussed later. Tension fractures normal to the direction of greatest stretching are radial in relation to the bending. In the gentle bends of a nearly horizontal stratum they are almost vertical.

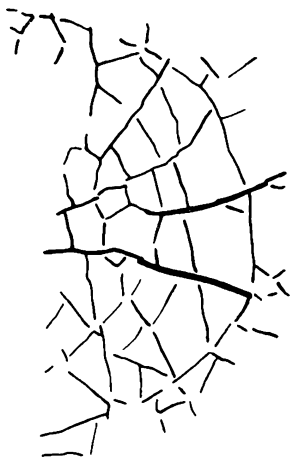


FIG. 263. Diagram of a turtle structure on a dome.

Ye. I. Chertkova's experiments have been discussed above; these have revealed that tension fractures depend on the shape of the uplifts. The same may be observed in the field. Experiments have shown that circular uplifts are accompanied by radial and concentric fractures, which have also been observed around circular domes in the field. If both fracture systems are developed, they produce a so-called turtle structure (Fig. 263). In some cases, however, only one of the fracture systems is developed, most frequently the radial. Figure 14 shows a case in which both systems are well developed: here the concentric fractures are filled by dikes. Concentric fractures are known near Khibinsk (in the Kola peninsula), where a domical uplift was created by a large intrusion. In elliptical uplifts (brachyanticlines), the orientation of the above fracture systems changes somewhat; here the radial fractures striking parallel to the dip of the beds and normal to

their strike are the most abundant. Radial fractures developed on both flanks of the uplifts frequently cut through the axes (Fig. 264).

Beside radial fractures, there are longitudinal fractures, most abundantly developed on the crests of uplifts. On the plunges of uplifts, combinations of longitudinal and radial fractures produce typical funnel-shaped structures.

Ye. N. Permyakov (1949), who has studied fractures in the rocks of the Russian platform, stated that longitudinal and diagonal fractures prevail in gentle uplifts. The direction of maximum bending to a certain degree determines the predominance of one system of fractures or another in the uplift: the fractures most commonly become oriented normal to the sharpest bending. The most intensive fracturing, as a rule, occurs in the crests of uplifts, where the rocks are most intensively squeezed, and gradually decreases toward the adjacent depressions.

Special study of this problem by M. V. Gzovskiy (1954) on models has shown that in the formation of a brachyanticlinal uplift as a result of vertical forces, longitudinal faults are the first to be produced. Their appearance changes the distribution of the stresses. The greatest tension is now along the axis of the brachyanticline, and therefore lateral fractures are formed after the longitudinal fractures.

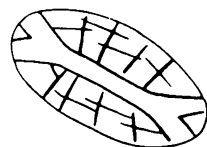


FIG. 264. Diagram of joints on an oval uplift.

Large anteclines are usually highly fractured; some fractures develop into cracks of great size; their position, in the ideal case, is controlled by the same geometric rules as in the case above.

In some small domes, and especially in broad anteclines, the large cracks frequently deviate from their regular directions because of the mechanical heterogeneity of the earth's crust and the existence of weakened zones at the contacts between different rocks. Bending fractures, for example, curve in the same directions as the borders of the intrusive bodies. Bending fractures also develop on massive uplifts in geosynclines, such as central uplifts or anticlinoria. The location of these fractures is not determined by the individual folds: groups of continuous folds are cut by fractures in directions determined by the general shape of the entire uplift as a whole.

The intensity of bending required to produce tension fractures depends on the properties of the material and on the conditions of the deformation, and in various cases it may vary widely. Even the relatively gentle uplifts of Tertiary and Quaternary age in eastern Siberia and in the Cordilleras of North America are intensively fractured. On the other hand, in a number of uplifts of the Swiss Alps no large fractures have thus far been discovered, and the bending of the earth's crust apparently took place in rocks that were more plastic.

An important factor in the development of faults in general, affecting

their frequency and direction, is the *fundamental divisibility of the earth's crust*—its block structure as determined by primary fractures. This will be discussed later.

It is widely believed that upwarping of the earth's crust causes more intensive stretching near the crust's surface than at depth. This is not quite correct. If the layers in a particular part of the earth's crust were originally flat-lying and were then upwarped by ascending pressure, the intensity of the stretching would be the same at depth and at the surface, provided the stratum contains beds that can slide past each other along the bedding planes. Stretching of the crust's surface and compression of its bottom occur only if the uplift is due to horizontal pressure. If the bending is caused by vertical forces, greater stretching can rather be expected in the depths. Domes formed of horizontal layers by ascending forces were mentioned above: it was noted that in

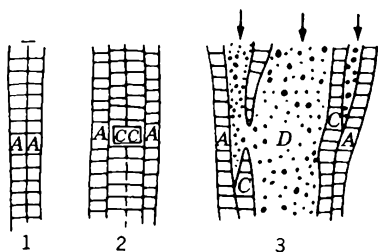


FIG. 265. Stages of reopening and filling of fractures: (A) first dolerite, (C) second dolerite, (D) quartz porphyry.

this case the lower beds are more sharply bent than the upper ones. Consequently the stretching is more intensive below than above (Fig. 258). Hence one might assume an earlier development of bending fractures at depth. The actual conditions are, however, more complicated, for the ambient pressure at depth increases the plasticity of the rocks and renders them capable of withstanding greater plastic deformation than at the surface of the crust.

The development of bending fractures at the crust's surface or in its depths depends upon three factors: the intensity of the bending, the overall ambient pressure, and the physical properties of rocks. Some domes are known in which open fractures cut through only the upper beds and die out at some depth. In other cases, the bending fractures are filled with magmatic material, indicating their development at depth and at least a temporary isolation from the surface. Geologic data sometimes permit us to establish the long life of tension fractures and their repeated reopening. This refers to complex dikes, which reveal that the cracks were formed as a result of several consecutive movements in which each opening was followed by a new intrusion of magma (Fig. 265).

Normal Faults

The stretching of the earth's crust during its uplift does not end with the formation of fractures. Because of the continued stretching, the tension fractures pass into faults, most commonly into normal faults.

As a rule, normal faults accompany the bending and stretching of the earth's crust, as may be seen from the particular features of their disposition. Normal faults rarely occur separately; they are usually combined in one way or another to form grabens and horsts or step faults. Grabens are the basic structure, accompanied by step faults along the flanks. Horsts are the secondary structural form and occur because certain parts of the earth's crust remain in place while the adjacent parts descend along the fault planes. Normal faults frequently occur along

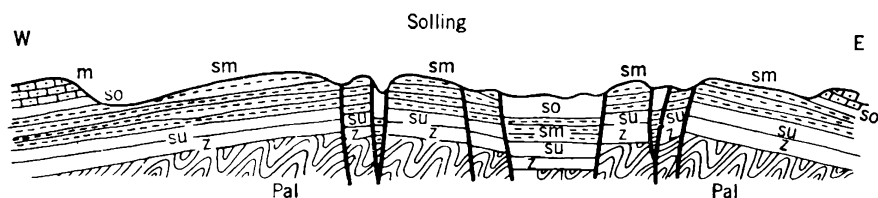


FIG. 266. Normal faults along tension fractures in the Solling uplift (Germany): Lower Paleozoic unconformably overlain by Permian sediments.

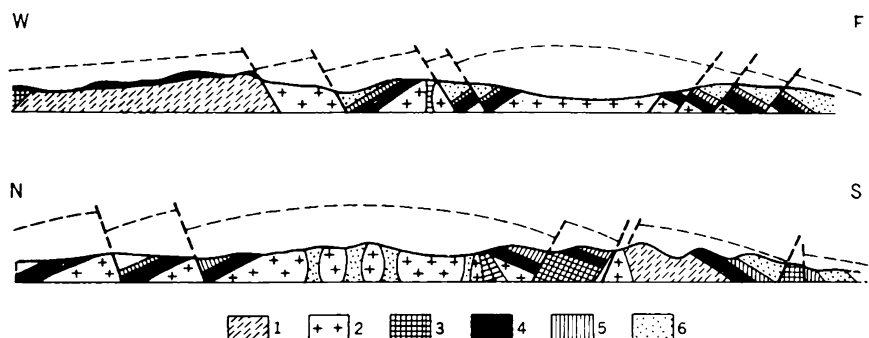


FIG. 267. Normal faults along shear fractures in the Morvan uplift (France): (1) gneiss; (2) granite; (3) Paleozoic; (4) Triassic; (5) Lias; (6) Middle and Upper Jurassic.

tension fractures. In this case, some part of the uplifted stratum terminated by tension fractures separates from the adjacent parts and is dropped to form a graben as a result of the continued stretching. In dropping, it wedges the adjacent blocks together.

Normal faults occur even more frequently along shear fractures. The faults developed on the basis of tension fractures are vertical or very steep, whereas faults developed on the basis of shear fractures have a smaller angle of dip (Figs. 266 and 267). Some faults may be formed along the combined plane of tension and shear fracture systems and, owing to this combination, may have different strikes and dips in various parts, each such part corresponding to one of the fracture systems entering into the fault. The normal faults along both tension

and shear fractures increase the area of that part of the crust's surface. This becomes evident if we take a cross section and return the vertically displaced blocks to their original position. The blocks cannot touch each other, and open spaces remain between them. The width of the open spaces is a measure of the extent of stretching.

Whereas normal faults related to tension are brittle, those of shearing are closely related to plastic deformation and are ductile. Nevertheless, only the spatial relation between faults and plastic deformation is accessible to direct geologic observation, whereas the relationship in time can only be visualized on the basis of indirect observations. The spatial relationship may take the form of a transition from normal faults of shearing to plastic deformation along the strike or dip. In tracing a fault plane toward its ends, one can see a gradual decrease in the displacement until it disappears into the fracture. Farther on, only small

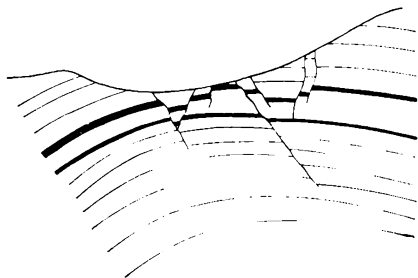


FIG. 268. System of normal faults in a domical uplift; the faults die out at depth. Apsheron peninsula.

slips along hardly visible fractures cause a gradual bending of the stratum, and finally nothing but plastic deformation remains. The known relationship of shearing to plastic deformation and the mechanics of concentrated displacements along a shear plane suggest that a fault passes into plastic deformation in both time and space: the end of the fault plane, where concentrated displacement has just begun, is its youngest part. On

the other hand, the places with the greatest displacement are the relatively oldest parts of the fault plane, where the process is far advanced. Thus a fault plane develops from the point where the largest displacement is observed to the places where there is minimum displacement.

If a normal fault dies out in depth, as in Fig. 268, the displacement may also decrease because of the reduced thickness of the lowered blocks. This indicates the beginning of faulting in a plastic stretching of the stratum, with a local reduction of its thickness in an initial neck. Experiments easily reproduce this phenomenon (Fig. 269).

Since the normal faults occur as a result of the further development of bending fractures, they retain the latter's relationship to other structural forms. Like bending fractures, normal faults occur within uplifts produced by vertical forces. Their positions are specifically related to the shapes of the uplifts involved and to the properties of the displaced material. Let us review some normal faults that are developed in large and small uplifts. Figure 270 shows a domical fold in which the tension

fractures pass into normal faults of relatively small displacement. Figure 266 shows a cross section of a small uplift disturbed by faults that have followed the lines of tension fractures, and Fig. 267 is a cross section of a dome cut by shear planes along which the hanging walls have dropped. This is the most typical form of normal faults that complicate

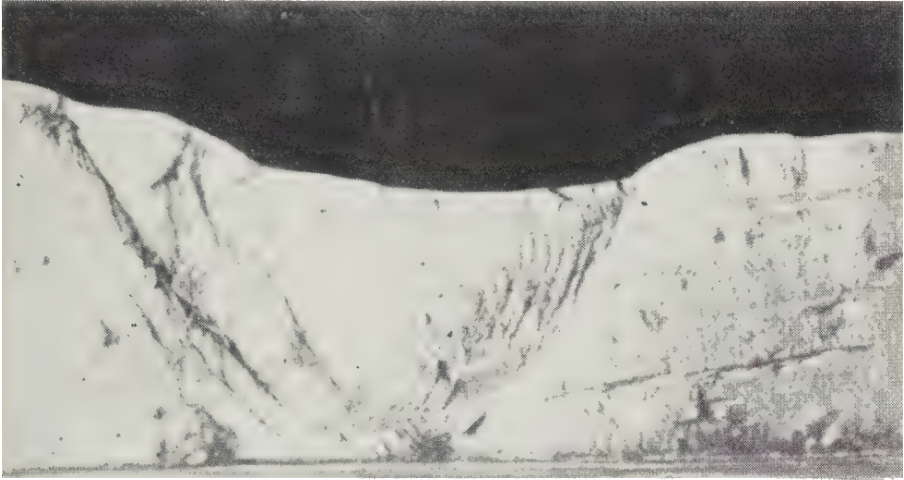


FIG. 269. Normal faults reproduced experimentally (*after Cloos*). On the right side the displacement along a large fault decreases with depth.

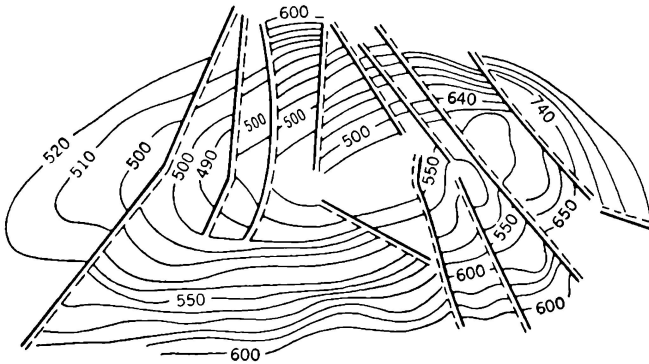


FIG. 270. An oval domical uplift cut by normal faults. Dashed lines indicate down-thrown sides; contours in meters.

the structure of an uplift: the closer the fault blocks to the crest of the uplift, the greater their downward displacement. Thus the normal faults are formed in the direction opposite to that of the uplift's elevation.

Central Germany contains many so-called anticlinal horsts, which are usually oval uplifts. Normal faults in the flanks are very typical

of these uplifts. They dip toward the adjacent structural depressions and displace the foot walls upward. In this case, the uplifting and faulting work in harmony, both elevating the central core of the uplift in question. These faults die out at depth, where they grade into plastic deformation. Western European geologists who still consider horizontal compression of the earth's crust to be the principal cause of deformation meet with great difficulty in their attempts to explain the formation of these folds. Actually, plastic bending of deep-seated rocks requires compression if it is caused by horizontal forces, whereas the structures formed by normal faults can only be the result of tension. Consequently, both compression and stretching must be assumed to have occurred in

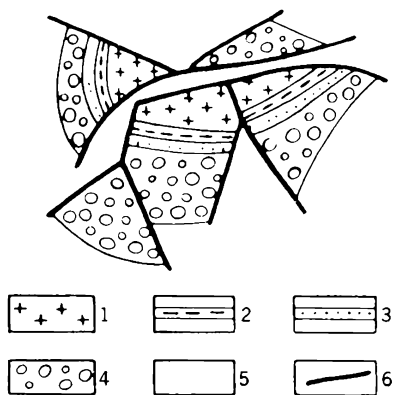


FIG. 271. Sketch map of the Shunat uplift in the Emba region: (1) Jurassic; (2-4) Lower Cretaceous; (5) Upper Cretaceous and Tertiary sediments; (6) faults.

the same place and at the same time in order to explain the anticlinal horsts. Western European geologists attempt to avoid this impossible condition by using an extremely complicated sequence of mechanical actions based upon combinations of differently oriented horizontal forces. The problem becomes simple, however, if it is assumed that the anticlinal horsts were formed not by horizontal but by vertical forces and that the plasticity of rocks increases with depth.

In this case, the vertical forces first produce the local oval uplifts by plastic bending; when the limits of plasticity are exceeded, cracking and faulting begin near the surface, while plastic deformation may continue at depth. Hence the stretching is first compensated by plastic deformation and then by normal faulting.

Interesting structures may be seen at the tops of diapir domes produced by pressure from their plastic cores. The stretching causes fracturing and faulting in the overlying stratum. The positions of the fractures and faults usually depend on the shapes of the respective domes. Since diapir domes usually have oval shapes, longitudinal and radial faults occur in combination. The middle of the dome is usually lowered, forming a central graben, while the flanks cut by the radial tension fractures are broken into separate blocks displaced vertically relative to each other (Fig. 271). The tops of some domes are so intensively broken up that they recall broken saucers. Figure 272 shows a cross section of a dome with a central graben: the lower parts of the upper complex are broken by tension into blocks separated from each other, and similar blocks from the upper parts of the same complex are

dropped into the fractures to form grabens (Borisov and Buyalov, 1938; Ruzhentsev, 1930; Shumilin, 1933).

As a result of the tension, fracturing, and faulting of very large uplifts, tremendous displacements have occurred in the Transbaykal region, the Rhine basin, and East Africa. According to the most recent studies (Pavlovskiy, 1948; Dumitrashko, 1948), the Baykal system of faults has produced a large number of grabens, including (besides the Baykal graben proper) the Tunka, the Barguzin, the Upper Angara, the Muya, the Toksimsk, the Upper Chara, the Bauntovsk, and many smaller grabens. All of them trend southwest to northeast, parallel to the axis of the

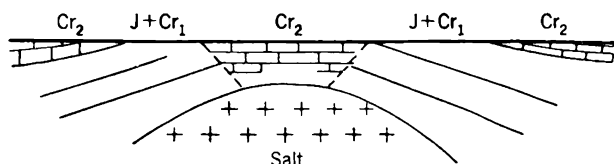


FIG. 272. Cross section of a salt dome with a central graben.

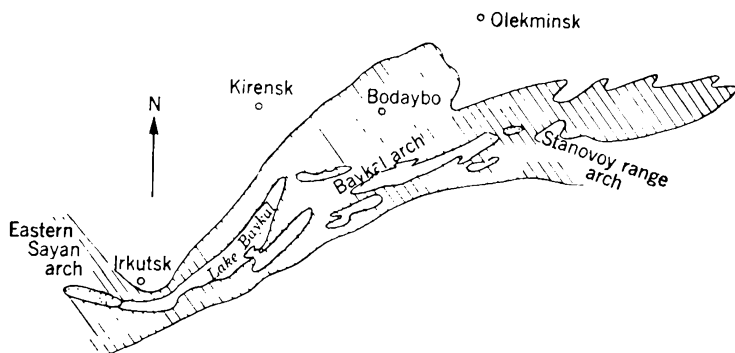


FIG. 273. Sketch map of the Baykal arch (after Ye. V. Pavlovskiy). Dotted areas are depressions along normal faults.

broad and gentle Baykal upwarp (Fig. 273), formed as an antecline within the platform. The upwarping began long ago, at least before the Jurassic period. As a result of further elevation and stretching during the Tertiary and Quaternary periods, longitudinal fractures developed, in some places earlier and in others later, and were followed by normal faults along which parts of the uplift descended to form grabens. The deepest of these grabens, the Baykal graben, is far below sea level, and the total vertical displacement along the fault planes bordering the graben apparently exceeds 2 km.

The graben system in the Rhine basin includes the Upper Rhine, the Lower Rhine, and the Hesse grabens (Cloos, 1939; Haug, 1932). The Upper Rhine graben separates the structural uplifts of the Black Forest

and the Vosges from each other, and the two other grabens cut through the Rheinische Schiefergebirge (Fig. 274). These uplifted areas are anteklises within the Alpine platform. Precambrian and Paleozoic rocks crop out at their cores; the margins are occupied by Mesozoic deposits. Paleogeographic data reveal that at the beginning of the Mesozoic, all these elevated areas were part of a single uplift, whose rise continued throughout the Mesozoic and the Early Paleogene. In the course of this rise, the rocks were subjected to stretching and fracturing; in the Oligocene the central core of the uplift sank, forming the Upper

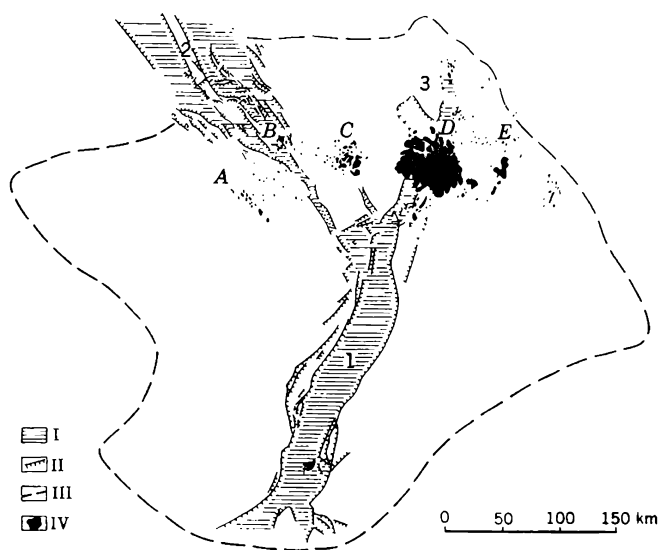


FIG. 274. Sketch map of the Rhine graben system: (I) grabens: (1) Upper Rhine, (2) Lower Rhine, (3) Hesse; (II) normal faults; (III) outlines of the Rhine anteklise; (IV) regions of Tertiary and Quaternary volcanic activity; (A) Eifel; (B) Neuwieder and Siebengebirge; (C) Westerwald; (D) Vogelsberg; (E) Rhoen.

Rhine graben and separating the Black Forest from the Vosges Mountains. The Hesse and Upper Rhine grabens were formed at about the same time.

The remaining parts of the uplift continued to rise, and the continued tension was compensated by still further sinking of the grabens. Owing to these continuous movements, the vertical displacement along the fault planes locally exceeds 1 km. The grabens are bordered not by single faults but by a system of fault planes, so that the above total displacement is the sum of the partial displacements along each fault plane.

The relationship of the grabens to the shapes of the respective uplifts is very interesting, and it reaffirms the general rule mentioned above. Figure 274 shows the grabens and the outlines of the entire upwarped

area. The strikes of the Upper Rhine graben follow the axis of the uplift and appear to be the trunk of the graben system. Their position is controlled by the bending fractures that trend parallel to strike of the stratum. Bending fractures parallel to the strike are typical, as we know, of the central parts of elongated uplifts everywhere. At both its north and south ends the uplift plunges, and the single graben splits into two branches. This branching follows the same rule mentioned above in discussing faults in diapir domes.

Still larger faults occur in East Africa (Cloos, 1939). Here there are two systems of faults: the East African system runs from the mouth of the Zambezi River toward Ethiopia through all of eastern Africa, passing through the region of the great African lakes; the graben of Eritrea lies between the Arabian Peninsula and Africa along the Red Sea. Both these systems meet in Ethiopia and form a tremendous complex of faults that occupy a meridionally elongated area 6,500 km long (Fig. 275). These graben systems are developed within anteklises formed during the Alpine tectonic cycle. The East African anteklise is located between the Congo River basin and the coast of East Africa. It is 2,500 km long and 1,500 km wide. Its central part has a stratigraphic elevation of more than 1 km relative to the margins. The elevation took place during the entire Mesozoic era and the Early Tertiary period. In the Oligocene the central part of the anteklise was broken by the two zones of the graben. The western zone of faults circles the western shore of Lake Victoria, and its further extension includes the basins of Lakes Rudolph, Albert, Tanganyika, and Nyasa. The second zone passes around the eastern shore of Lake Victoria, and its northward extension meets the first zone near Lakes Rudolph and Nyasa.

The grabens are reflected topographically as deep, broad valleys with steep slopes and a chain of lakes at the bottom. The bottoms of some lakes are below sea level (the bottom of Lake Tanganyika is 500 m below), whereas the water tables of the lakes are 770 m or more above sea level. The vertical displacements locally reach 2 km. Individual faults do not extend along the entire zone, but die out somewhere within it and are replaced by new ones, offset to a certain distance, which then continue along the previous strike.

The graben system of Eritrea, which includes the grabens of the Red Sea, the Gulf of Suez, Syria, northern Ethiopia, and the Gulf of Aden, was formed in the central part of another anteklise but at the same time as the system described above (Fig. 275). The vertical displacement in places exceeds 2 km. To this graben system belongs one of the deepest depressions on the earth: the Dead Sea basin, of which the water table is nearly 400 m below sea level. The relationship between the grabens and the shapes of the uplifts is similar to that of the Rhine graben system. The graben system of Eritrea consists of a central

trunk—the Red Sea graben, which follows the axis of the antecline—and its branches located at both ends of the uplifts. These branches are the grabens of the Gulf of Suez and Syria in the north and those of the Gulf of Aden and Ethiopia in the south.

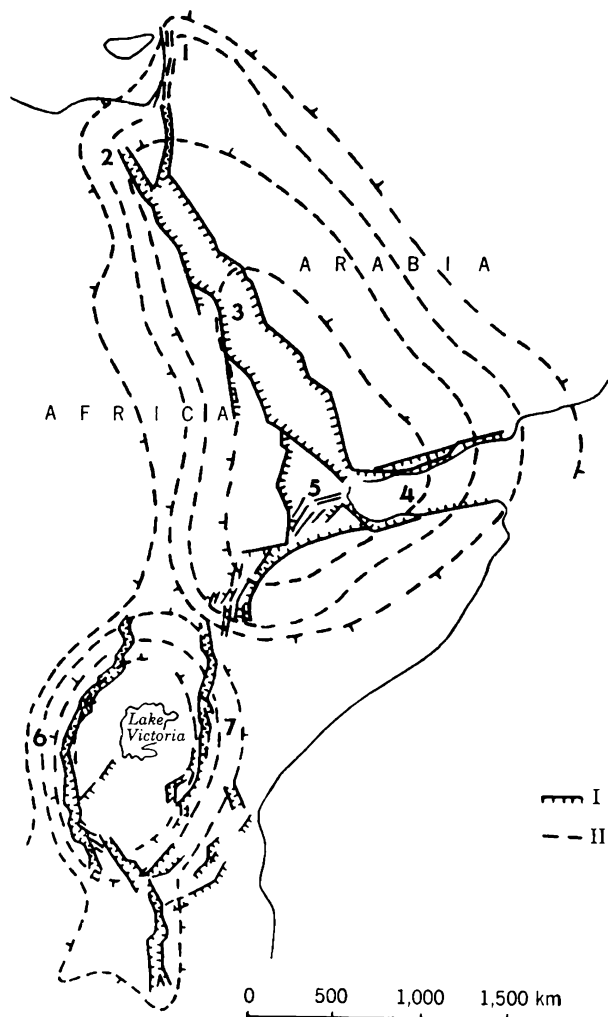


FIG. 275. The African graben system: (I) normal faults; (II) structure contours of uplifts (anteclines); (1-7) grabens; (1) Dead Sea graben, (2) Suez graben, (3) Red Sea graben, (4) Aden graben, (5) Ethiopia graben, (6) East Africa (western belt) graben, (7) East Africa (eastern belt) graben.

Figure 276 compares the graben systems of the Rhine and Eritrea. The Rhine grabens are shown on a scale ten times larger than those of Eritrea, in order to make the comparison easier; for the same reason, the Rhine system has been turned 180°. Comparison of the two regions

clearly demonstrates the similarity of the mechanical laws of faulting, regardless of the scale on which it occurs.

The East African graben system also trends along the axis of the antecline with which it is associated. Here, however, there are some complications: the branching of the graben system in its central part into two zones that pass around Lake Victoria on both sides. This complication appears to be due to the local mechanical conditions: the area of Lake Victoria is occupied by a large oval granitic batholith surrounded by gneisses and shales whose cleavage and schistosity are vertical. The tension in this case produced fractures at the batholith's contact and in its ring of schistose rocks.

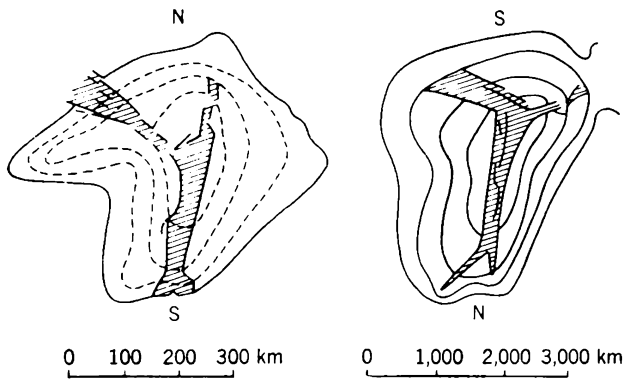


FIG. 276. Comparison of the Rhine and Eritrea grabens. The graben system of Eritrea (right) is shown at nearly one-tenth the scale of the Rhine system. Contour lines outline the anteclines.

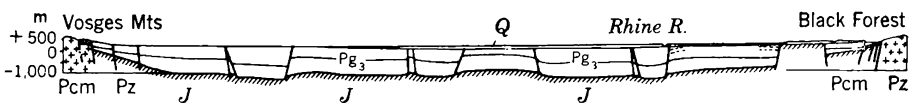


FIG. 277. A complex graben. Cross section through the Upper Rhine graben.

In concluding this survey of normal faults, some common combinations of faults may be mentioned. It has already been pointed out that faults constantly occur together and form groups. Large graben systems are always, therefore, a combination of grabens and horsts, large and small, contained within each other, and only the total displacement produces the general depression (Fig. 277).

Figure 278 shows the so-called *Y* faults, which are very common. Here one may see a major normal fault and a system of smaller faults branching from it like feathers into the hanging wall, forming a pattern resembling the letter *Y*. The system of feather fractures appears to be caused by friction: slip along the major fault produces shear deformation because of the friction in the fault plane, and the shear, in turn,

causes tension fractures normal to the maximum tension. Further slip along the major fault, or the friction due to this movement, bends the layers of the hanging wall and displaces them along the tension fractures. As a result, as seen in Fig. 278, the blocks turn and are displaced toward the major fault. If outcrops are rare, the rotation of the beds

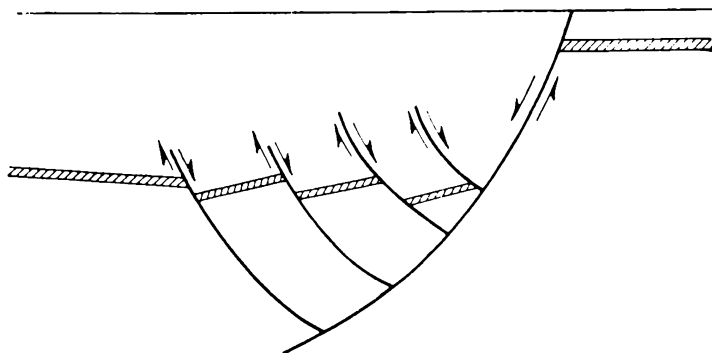


FIG. 278. Y-shaped faults.

in the fault zone may be interpreted erroneously as folds in areas of flat-lying sediments.

There also are X faults, which occur as a result of the intersection of inclined faults (Fig. 279). The displacement along these faults occurs at different times and produces a combined result in which the upper part of the structure appears to be a graben and the lower part a horst.

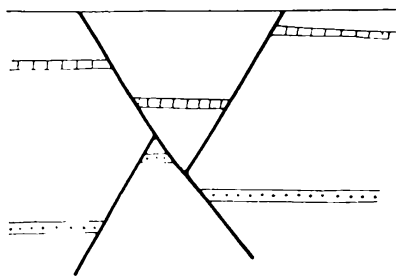


FIG. 279. X-shaped faults.

Normal faults also have various horizontal combinations. They may cut or terminate each other at various angles. Combinations of longitudinal and transverse faults are very abundant. The depressed or elevated blocks are sometimes bordered on all sides by faults (forming oval depressions and elevated blocks). One-sided grabens or horsts may also acquire various shapes as the faults die out along their strike. Grabens shaped like

a hammock and horsts in the shape of bridges may occur for the same reason. A change in the direction of the vertical displacement along the fault plane may produce so-called hinge faults (Fig. 280).

General Jointing

In the preceding chapter, tension and its associated fractures were described. This fracturing can result not only from external tectonic

forces but also from internal tensions due to contraction, such as that caused by cooling, desiccation, or physicochemical changes in the rocks involved. These circumstances produce tension fractures that may involve the entire area and form general jointing. The position and orientation of these fractures are usually predetermined by the anisotropy of the rocks, caused by preceding deformations or the orientation of the minerals. These fractures, as far as they are presently known, cannot always be distinguished from tension fractures of tectonic origin. The two groups of tension fractures can easily be confused, since they are produced by the same kind of stress; and the orientation of contraction fractures, depending on the anisotropy of tectonic origin, may coincide with that of the tectonic fractures. For example, the two groups of tension fractures can hardly be distinguished at the tops of domes, antecises, and other uplifts: the radial, longitudinal, and concentric fractures formed in such uplifts may be of tectonic origin (directly related to the stretching caused by tectonic forces), but they may also be caused by internal tension resulting from contraction. The latter's orientation will depend on the anisotropy produced during the plastic deformation of the stratum.

In other cases, the general jointing is more easily recognizable. This category includes fractures developed in undisturbed flat-lying areas on platforms. They are also found in undeformed lava flows. Even in folded beds contraction fractures may be distinguished: here they occur, it may be assumed, in the form of widely developed fractures within individual beds, and they are oriented normal to the bedding planes. These fractures remain closed even in the crests of folds. If they were bending fractures formed by tension during folding, they would have been open in the crests of the folds, where the beds are sharply bent. If they remain closed, they must evidently be regarded as fractures formed through contraction of the rocks and oriented according to the anisotropy resulting from the plastic deformation of an earlier period of folding. This view must, however, be considered to be an assumption rather than a final conclusion.

Within intrusive masses enclosed in folded strata, contraction and

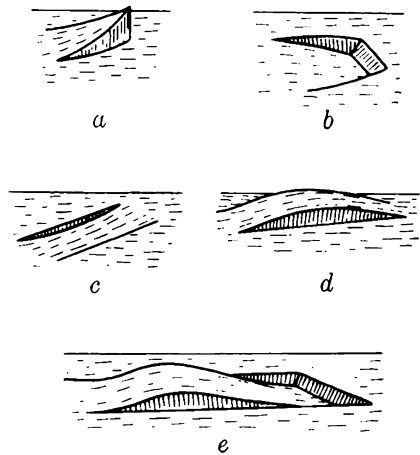


FIG. 280. Various types of normal faults in which the amount of displacement changes along the strike: (a) one-sided horst; (b) one-sided graben; (c) cradle-shaped graben; (d) bridge-shaped horst; (e) hinge faults.

tectonic fractures are also very hard to distinguish. Since the intrusives have undergone tectonic deformation, it is very likely that they contain fractures of tectonic origin. But the present author believes that direct tectonic action as the cause of fracturing in the intrusives is usually exaggerated. It is more likely that tectonic forces affect the fracturing in intrusives only indirectly: as in the previous case, the tectonic forces do not directly produce fractures, but they affect the orientation of the contraction fractures, which orient themselves in accordance with the anisotropy created when the magma was still not solidified. This concept appears to be well founded, since the intrusives reveal a clear relationship between fracture systems and flow structures.

Some examples of the disposition of contraction fractures may be cited, with the reservation that the problem is still inadequately studied. In the first place, we may discuss the disposition of tension fractures in a folded stratum. The majority of contraction fractures are oriented normal to the bedding. According to their strike, they can be divided into several systems. If it is imagined that the beds are returned to their original horizontal position, three systems of fractures will appear to prevail: (1) a system whose strike is parallel to the dip of the beds after folding; (2) a system whose strike is the same as that of the beds; (3) two systems whose strikes are at a 45° angle to the strike and dip of the stratum (Kirillova, 1949; Semenenko, 1946). If the stratum changes its strike because of the shape of the fold, all the fractures are rotated to a certain degree. This demonstrates that unreliable data can be obtained if, in striving to obtain the greatest possible number of statistical measurements, one records the strikes and dips without paying attention to the particular part of the fold in which the fractures occur. Among general jointing fractures there are also some with orientations different from the above systems; these, however, are less abundant.

The relationship of the contraction fractures to certain beds reveals that the anisotropy, which affects the orientation of the fractures and is in turn a result of plastic deformation, is slightly different in each particular bed. The orientation of the plastic movement depends on the shape of the folds. The general jointing, therefore, depends on the strike and dip of the particular stratum concerned. It is still questionable, however, which specific textures reflect the anisotropy, since exact microscopic studies in this regard have not yet been undertaken. Such studies might provide extremely useful data for understanding certain peculiarities of the movement of the material in the beds during folding.

As long as the rocks continue to decrease in volume and contraction fractures continue to form, the beds must inevitably slide past each other along the bedding planes. We know that the friction produced in this sliding creates several centers of strain and breaks the rocks in a number of fractures. The extension of the latter depends, first, on the

intensity of the internal stress and, second, on the amount of friction in the bedding planes. The greater the friction, the larger the fractures that must be developed, since a large amount of friction can be overcome only if the strain is integrated within larger blocks. Thus it becomes clear why the fractures are less abundant and extend farther in the depths of the rocks. Observations made in underground workings reveal a decrease in the number of the fractures and an increase in their extent. It is clear that the smaller frequency of fractures is related to the increasingly greater pressure at depth, which, in turn, causes greater friction between the beds.

Thus the problem of the formation of contraction fractures in folded strata is still very unclear. The situation is even worse, however, in the case of flat-lying beds on platforms. As far as they have been observed, the contraction fractures in flat-lying beds have definite orientations: they are vertical, and each system maintains its strike within a certain area. The contraction fractures within the Russian platform were studied by N. S. Shatskiy (1945), who believed that the entire platform has a single pattern of fractures. According to other sources, the Cambrian and Silurian deposits in Estonia and Gotland are everywhere cut by fractures, among which equatorial, meridional, and diagonal systems are especially well developed. Ye. N. Permyakov (1949), whose work has been mentioned earlier, believed that all the fractures on platforms are of tectonic origin and related their disposition to the outlines of tectonic elements of different extents. A. S. Novikova (1951), however, considered all the joint fractures on platforms to be the result of factors related not to tectonics but to the contraction of rocks during their consolidation, and she found no clear regularities in their deposition. The factual data are still very limited, and the problem is still a subject for further study. The suggestion has recently been made that the directions of general fractures on platforms are in some manner associated with the fundamental divisibility of the earth's crust (see below).

Let us discuss the contraction fractures that occur in intrusives (Balk, 1937; Pek, 1940; Polkanov, 1934). It has been said that the disposition of the fractures in intrusives (see Fig. 41) is closely controlled by the flow structures. Here the relationship between the anisotropy of the rock and the tectonic deformation is evident; this favorably distinguishes igneous rocks from sedimentary strata, in which the changes in texture due to plastic deformation are not always recognizable.

The fractures normal to the flow lines are the most important in intrusive bodies. They are usually most clearly pronounced in the marginal parts of an intrusive and frequently disappear entirely within its center. But the fractures may extend even to the centers of intrusives, thereby indicating orientation of the minerals, at least in barely recognizable form, as far as the core of the intrusives. Such normal fractures

develop in the early stages of solidification and shrinkage of the intrusives and are consequently filled with dikes of aplite, pegmatite, and other products of igneous activity.

The total width of the fractures developed across the flow lines within certain parts of intrusives reveals the great shrinkage of the rocks during their cooling period (an average of 4 to 5 per cent). A greater decrease in volume apparently occurs when the molten magma solidifies, thereby loosing a great amount of its volatile components.

The next group of fractures in intrusives includes those oriented parallel to the flow lines. These longitudinal fractures are also developed before the magma has completely solidified and frequently contain dikes of aplite, pegmatite, quartz, and basic rocks. Observations reveal that among the fractures of this group two types can be distinguished. The first type (longitudinal fractures proper) includes vertical or very steep fractures; the other type consists of fractures with gentle dips developed near the top of the intrusive, where the flow lines are also nearly horizontal, whereas near the steep contacts they dip steeply. In other words, fractures of the second type are more or less parallel to the contacts wherever the flow lines are also parallel to the contacts. The above fracture systems are formed during solidification and cooling as a result of contraction of the intrusive body, and their disposition is predetermined by the anisotropy of the internal structure of intrusive rocks. Briefly, these fracture systems break the intrusives into blocks shaped like parallelepipeds.

It has been mentioned that in uniformly crystallized intrusives, flow lines are absent or scarcely distinguishable and that, consequently, the fracture pattern loses its angular form and becomes concentric, egg-shaped, pillow-shaped, etc. The tension fractures caused by internal stresses include both fractures parallel to the earth's surface and weathering joints. If consolidation causes shrinkage through the loss of water content or cooling, the tension fractures formed at a certain depth do not entirely relieve the stresses in the rock; part of the stress is maintained, since the internal bonds between the particles of rock resist contraction. Near the surface, the rock undergoes physicochemical decomposition and, consequently, the above bonds are gradually broken. This is a possible reason why internal tension due to contraction may cause intensive fracturing near the surface in the zone of weathering. If this assumption is correct, the near-surface weathering fractures must retain the orientation of the older fractures; only their frequency will become greater than it is at depth. Thus the orientation of the surficial fractures may reveal that of deeper fractures. But the data upon which this concept can be based are still inadequate. The behavior of rocks in the zone of weathering more clearly reveals stress-related properties that are less pronounced in the rocks at depth.

COMPRESSION

Horizontal compression causes plastic elongation in other directions, since plastic deformation, as we know, does not change the volume. A reservation was made that, in the preliminary approach to some aspects of tectonics, deformations would be discussed in two dimensions. Therefore, in our constructions, compression along one dimension causes elongation along the other. Fracturing due to compression occurs first of all along the shear planes. Displacement along the cracks continues the plastic deformation and leads to further shortening along one axis and elongation along the other. Shear fractures are usually developed in groups, and they are combined in such a way that all of them in effect lead to shortening in one direction and elongation in another.

The orientation of the axis of deformation may differ. If the compression is horizontal, the elongation may be horizontal or vertical, depending on the direction of less resistance. If the compression is vertical, the elongation may be in any horizontal direction. But the deformation axes may also be inclined at various angles. Shear planes may be developed into overthrusts, strike-slip faults, or faults that combine both vertical and horizontal displacements (strike-slip thrust faults and overthrust strike-slip faults). Figure 281 shows a cross section of a group of thrust faults. Each displaces the rocks in such a way that their combined result is a horizontal shortening of the particular part of the earth's crust and an increase in its vertical dimensions. In bedded sedimentary strata, compression first of all produces folds formed by the plastic deformation. Faults must therefore be considered together with fold forms, using the latter in every particular case to determine the general features of the deformation and the orientation of its shortened and elongated axes.

It is known that folding causes a decrease in the thicknesses of the beds in the flanks of the folds and an increase in the crests and troughs. Because of the differing plasticity of rocks, some beds may be squeezed toward the crest more intensively than others. Consequently, the material moves toward the crest within each bed independently of others. But the heterogeneity or homogeneity of physical properties is not a changeless quality of a given stratum. A stratum originally composed of layers with widely different physical properties may become more or less uniform in the course of the deformation. Hence the rocks that

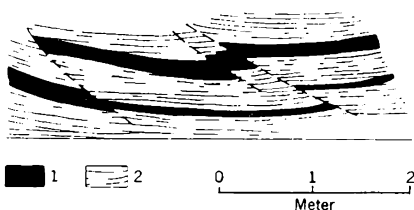


FIG. 281. Small "laminar" conjugate thrust faults. Paleogene. Trialet region (after I. V. Kirillova): (1) sandstones; (2) shales.

underwent deformation in different ways at the beginning of the compression may later, because of the advanced plastic deformation, become nearly uniform (Kirillova, 1949). The change in physical properties in the course of the deformation leads to the formation of very complicated structures in which the earlier interlayer displacements of the substance are overlapped by structural forms that involve the entire stratum uniformly as a whole. On the other hand, within a uniformly deformed stratum secondary or minor deformations can be distinguished, which one way or another are combined with the larger structural forms. The squeezing of the material from the flanks toward the crests of folds never proceeds evenly. The flow of material within certain beds may be much faster than in others. Moreover, even a single bed may reveal varying rates and intensities of the plastic movement in different parts. Thus certain parts of beds may be squeezed more than others and the displaced material concentrated somewhere else along the same bed, causing a local thickening. Hence the flanks of folds, where the beds generally are thinner, may have some zones of intensive squeezing, and other zones have become thickened. This has been discussed earlier.

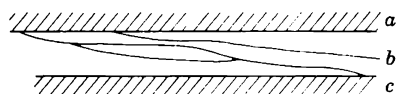


FIG. 282. Intraformational ruptures (after Ye. Ye. Zakharov): (a and c) sandstones; (b) shales.

Plastic deformation, which causes greater compression of the rocks in the flanks of folds, may under certain circumstances pass into faulting. Here the development of the fault planes begins with fractures that first cut through one and then a number of beds and finally through the entire

stratum, resulting in displacements. The types of such faults and their combinations depend on their location within the squeezed or thickened zone of the fold's flank.

Figure 282 shows fractures within a bed of shale enclosed between two sandstone beds (Zakharov and Korolev, 1940). In these fractures one will recognize shearing caused by compression of the shale between the sandstones and the extension of the shale bed parallel to the bedding. Since the fractures do not extend beyond a particular bed, it is difficult to say whether they occur within a zone of squeezing or of concentration of the plastic material. Figure 283 shows various shear fractures formed under similar conditions. The fractures extend, however, beyond a particular bed (Usov, 1926, 1933). All these fractures are located within thicker zones of concentration of the material—in those zones in the flanks where certain parts of the material, moving plastically toward the crests of folds, overtake and overlap the adjacent parts. In Fig. 283a we see another case in which the shear fractures extend beyond a particular bed and partially cut the adjacent beds. As a result of the uneven upward squeezing of the beds, the lower parts of

the beds overtake and overlap the upper and produce small platelike thrust faults. Another case of local concentration of the plastically moving material and the overtaking of certain flows by others is shown in Fig. 283*b*. The middle bed, apparently because of greater compression

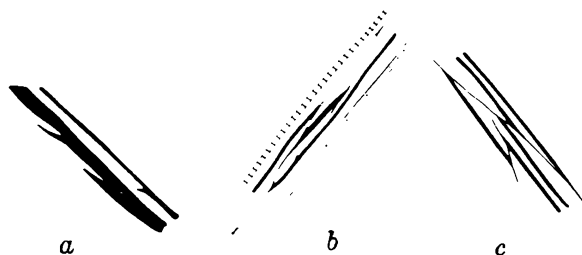


FIG. 283. Small faults extending beyond a single bed (after M. A. Usov).

somewhere deeper along the dip, came to be extended farther than the adjacent beds and consequently was broken into lenses, which, in moving upward, overlapped each other three times. This type of displacement should be called a *push fault*, since the movement undoubtedly was initiated in the lower parts of the flanks and pushed the higher ones, causing the dislocations in them.

Figure 283*c* shows another similar case, with complications: here the overtaking of the upper lenses by the lower ones produces one thrust fault (the lower fracture) and one push fault (the upper fracture). As a result, the lower lens is pushed between the two upper lenses.

If similar displacements occur on a greater scale, they produce large faults located in the flanks of folds (Rumyantsev, 1928). These thrust faults in one or both flanks of an anticline may move the flanks toward the crest of the fold (Fig. 284*a*). Displacements

toward the axes of synclines may appear in similar forms. Faults may be developed on one or both sides of the fold axes and move the trough of the syncline toward its flanks (Fig. 284*b*). Here the term “push fault” could be applied again, for the original movement is directed from the flanks toward the trough of the syncline.

The zones of accumulation of the material, where portions of the

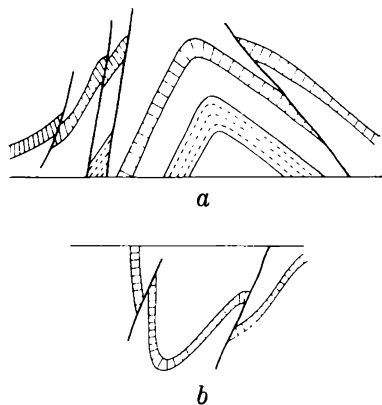


FIG. 284. Overthrusting of the flanks of an anticline over its crest (*a*) and underthrusting of the flanks of a syncline beneath its trough (*b*). Kuznets basin (after S. Rumyantsev).

same bed may overlap each other several times, must have corresponding zones of intensive squeezing from which the zones of concentration obtain the additional material. The entire flank, however, may be such a zone of squeezing, since squeezing is the principal deformation in the flanks of folds. If there is no particular zone of extremely uneven movement—no zone in which lenses of the same bed overlap each other twice or more—then the entire flank may be considered a single zone of squeezing.

Overall and zonal squeezing as a result of slip may occur along a single system of fractures or conjugate groups. Displacement along a group of fractures leads to a decrease in the thickness of the beds involved and their extension along the dip. Figure 285 shows various cases of such fault structures. Figure 285a is a picture of an overthrust along a single fracture: here squeezing leads to extension of the bed. In previous cases there was doubling and tripling of beds, but here we

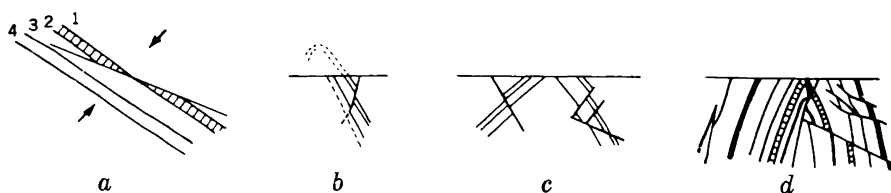


FIG. 285. Stretching of the flanks of folds, taking the form of fault systems (after A. Belitskiy).

see the contrary: an open space is formed between the two separate parts of one bed. A single thrust fault is also shown in Figure 285b. The next two illustrations show more complicated structures that result from combinations of fractures. In general, thrust faults occur along these fractures, but in some cases it would be better to call them under-thrusts. Their combination leads to contraction of the layers normal to the bedding planes and to extension parallel to the dip. The original occurrence of the layers prior to the fracturing must be reconstructed in order to draw a diagram of the deformation. In so doing, one can see that in these cases the displacement along the faults appears to be due to the different degrees of squeezing in different parts of the beds involved. In similar cases some authors speak of *packing*.

Vertical extension due to fracturing may sometimes be observed in both flanks of folds (Fig. 285c). In other cases, however, squeezing and overlapping of the beds may take place in one of the flanks of the fold and spaces may be opened between the broken parts of beds in another. This combination represents one type of deformation in one of the flanks and another kind in the other. Very impressive examples of differ-

ent deformations in different flanks can be observed in the Kuznets basin. The folds of this basin, whose trends are approximately meridional, have sharply asymmetrical flanks: the western flanks are characterized mostly by doubling and overlapping, whereas on the eastern flanks elongation along the dip and opening of spaces between the broken parts of beds are most common. This asymmetry of the flanks usually is closely related to overturned folds. In the overturned steep flanks of folds, as a rule, elongation along the dip is most typical, whereas in the gentle flanks overlapping of beds prevails. This kind of asymmetry of folds is very common in the Kuznets basin and appears to be closely related to the unequal value of the shear fractures. In all the above cases, the faulting, which causes squeezing or overlapping of the beds, always occurs along the dip. But dislocations of the material may also occur along the strike, including large strike-slip faults. In the same direction the beds may be overlapped. The fractures in this case are strike-slip faults and are diagonal to the directions of compression and elongation. Since horizontal compression causes elongation



FIG. 286. Thrust faults against the dip in vertical folds.

of the material along another horizontal axis less frequently than along the vertical axis, strike-slip faults are not as abundant as thrust faults. Displacements in the exact direction of the strike or dip are extremely rare; they have been considered because these extreme examples demonstrate the displacements most clearly. Nevertheless, displacements generally occur along neither the dip nor the strike but somewhere in between. Thus in nature one most commonly finds displacements of intermediate types, such as strike-slip thrust faults or thrust strike-slip faults.

Having reviewed displacements within particular beds or particular parts of folds, we may now discuss large faults, the regularities of which depend on the overall deformation of the entire stratum and on the folded structures in the area concerned. If a certain part of the earth's crust is compressed horizontally or subjected to tension vertically, or rather if the material has moved upward, thrust faults may be developed along the two systems of shear fractures (Fig. 286). Thus the folds continue to rise vertically, maintaining their horizontal symmetry.

In nature, however, such folds are rare, and usually one system or another of shear fractures develops more intensively. The reason for this is related to the nature of the horizontal compression, which causes

not exactly vertical elongation but elongation along a somewhat inclined axis. It is obvious that if the elongation is along an inclined axis, the fractures oriented more or less parallel to the elongation have a greater chance of grading into shear faults than the fractures with other orientations. Consequently, the shear fractures that dip in the same direction as the axial planes of the folds more frequently develop into faults. Displacements along these occur in the form of thrust faults. Actually, a slight inclination of the fold's axis suffices to suppress one system of shear fracture by the other, for as soon as the displacement begins along one system of the shear planes, the shifting load along the other disappears and no displacement can take place there.

Folds with vertical axial planes are found only in narrow folded zones, whereas in greater parts of the folded region the axial planes of the folds become inclined in one direction or another; usually folds with the same inclinations occupy broad zones. Thus the thrust faults accompanying the folds are oriented in a certain direction within each of these broad zones. Observations show that there are thrust faults the positions of which are related to folds, whereas other thrust faults cut through the folded stratum without being associated with a particular fold. The former are called *fold thrust faults*, and the latter are known as *intersecting thrust faults* (see Chap. 9).

Intersecting thrust faults are apparently formed in places where the dip of shear planes differs considerably from that of fold axial planes. As a rule, the faults are more gentle than the axial planes of the folds. Figure 72 shows a group of such thrust faults. A similar thrust fault can also be seen on the right of the cross section in Fig. 24. If the orientation of the shear planes is close to that of the axial plane, the thrust faults are related to the folds and are contained within certain parts of the latter. Thrust faults commonly occur in the overturned flanks of folds. If many similarly overturned folds are cut by thrust faults, the imbricate structure known in many folded areas develops.

The reason for the more extensive development of thrust faults in overturned flanks is perhaps related to the differing dips of the layers relative to the thrust faults. On the upright flanks of folds, the thrust faults are steeper than the bedding planes and the upward movement of the hanging walls must overcome greater friction, since the beds there are cut at sharp angles. In the overturned flanks the faults are less steeply inclined than the bedding planes and the displacement can take place with considerably less friction than in the former case, since the beds are readily overturned in the direction of displacement and lie on fault planes with their reversed sides. Consequently, it can be thought that the plastic displacement is to a greater extent associated with the overturned flanks of folds and gradually passes into a shear fault because of the smaller resistance.

The concentration of the plastic movement in the overturned flanks may be established by a comparison of the various structures in which one can observe different stages of structural development. It is possible to distinguish cases that will represent consecutive stages of the process. First are the overturned flanks in which the contraction of the layers does not differ from that of upright flanks; then the beds become stretched and increasingly squeezed, until they are finally cut by thrust faults. The plane of the thrust naturally does not occur along the entire length of the flank concerned but develops within a certain part of it and gradually extends both up and down the dip and laterally. Consequently one may encounter thrust faults of widely varying extent with displacements from a few millimeters to several kilometers. Each thrust fault, regardless of its size, does not extend indefinitely. It dies out in every direction, passing into plastic deformation. Low-angle thrust faults with displacements reaching several kilometers are usually called *overthrusts*. At the beginning of the present century geologists were widely attracted by the concept of tectonic nappes. Many believed that tectonic thrust sheets or nappes were the most typical structures in mountain regions. It was thought that the extent of the horizontal displacement of the tectonic thrust sheets could be as great as 200 km. Particularly great nappes were assumed to exist in the Alps, whose general structure was depicted as consisting of a number of nappes lying one above the other and displaced to great distances. At the same time it was thought that tectonic thrust sheets were formed from recumbent folds of enormous dimensions.

At the present time, there has been a radical change in the accepted view of the importance of tectonic nappes in the structure of mountain regions. More detailed and objective investigations have indicated that the role of nappes was formerly greatly exaggerated (Belousov, Gzovskiy, and Goryachev, 1951). It appears that nappes associated with recumbent folds actually do exist in nature, but the amplitude of their displacement is evidently nowhere more than 10 to 15 km. They are often connected with injection folds, such as, for example, those located above a small nappe in a plastic series in Provence (Fig. 185). There are also horizontal displacements of greater extent, but in the light of the most recent data these are connected not with recumbent folds but with the gravity sliding of individual blocks of the earth's crust down the slopes of mountain ranges. An example of such gravitational thrusts is shown in Fig. 287. In this case, the sliding was facilitated by the fact that at the bottom of the sliding part of the crust there was a gypsiferous Triassic stratum consisting of extremely plastic and "slippery" rocks. In the western Alps the plastic Trias, and the almost equally plastic Eocene flysch series, generally played the determining role in the formation of gravitational thrust sheets. The latter some-

times have the appearance of enormous swellings of plastic rocks (Fig. 288).

The horizontal amplitude of the displacement of these gravitational thrusts in some places reaches 40 km. It may be assumed that the great

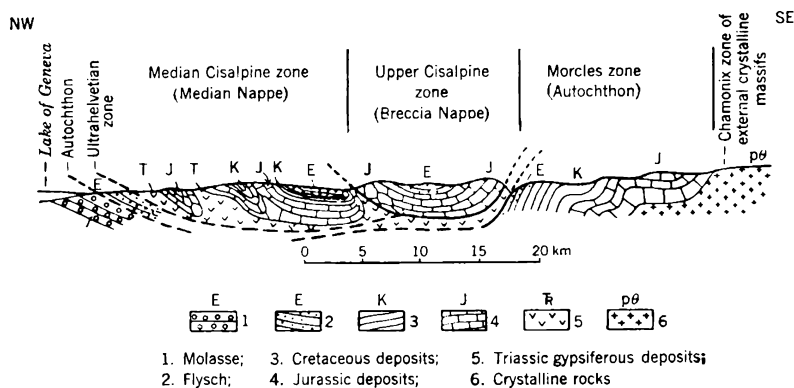


FIG. 287. Slide thrusts in the Alps. Diagrammatic section through the Upper Savoy Cisalpine region: (1) molasse, (2) flysch, (3) Cretaceous deposits, (4) Jurassic deposits, (5) Triassic gypsiferous deposits, (6) crystalline rocks.

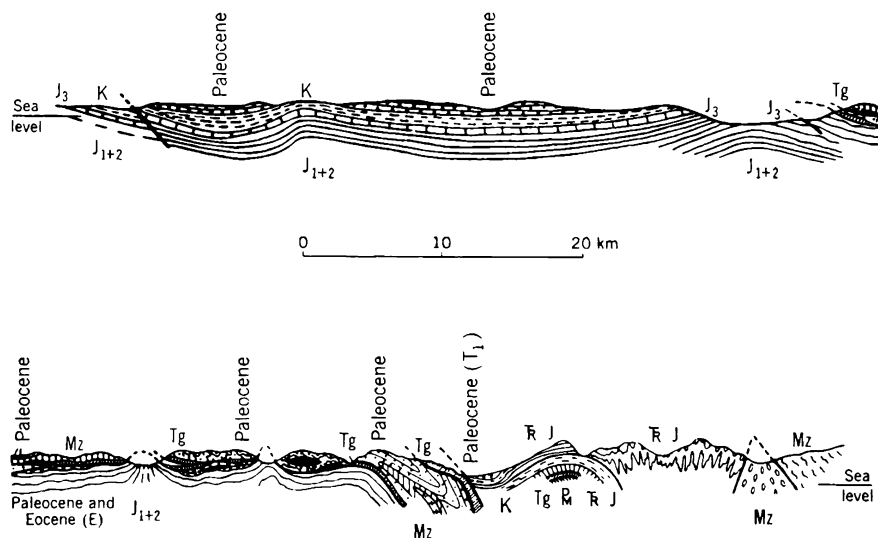


FIG. 288. Gravitational thrust in the Alps. *Upper*, diagrammatic section through the Durance River region. *Lower*, the Embrunais thrust sheet.

displacements did not take place all at once but gradually, by separate stages, as the rising mountain range grew and expanded and its slopes were displaced outward from the axis of the range. Such thrust sheets are completely separated from their "roots"—that is, from the complex

of rocks that remained in place—and there is no basis for thinking that at any time after the formation of the nappe its roots were in immediate contact with the body of the nappe. The mechanics of the formation of such thrusts are fully analogous to the mechanics of slide formation: certain blocks of the earth's crust were torn away and moved freely or flowed downward under the action of the force of gravity. The term *nappe vestige*, introduced earlier, is apparently not suited to thrust sheets of this type, which are primarily a combination of blocks torn from their roots and from each other and displaced along the slopes.

One may, of course, express doubt of the likelihood of formation of such huge "slides" on such gentle slopes as are usually found in mountain ranges. But here the phenomenon of creep undoubtedly plays a great part: plastic rocks lying for a long time on a slope will gradually and very slowly travel downward. This process can apparently take place on slopes of no more than a few degrees.

Gravitational thrust sheets are a phenomenon combining the properties of both surficial and tectonic processes: a gravitational thrust sheet resembles a slide, but its dimensions are such and it is accompanied by such additional phenomena (crumpling of the beds into very large folds, formation of breccia by friction, etc.) that it would be strange not to see in it some manifestations of the tectonic activity of the earth's crust.

On geologic maps, exposures of overthrusts appear in the form of wavy lines generally parallel to the folding. The wavy shape and the fact that they are parallel to the folding are the most important signs by which one may recognize overthrusts among strike-slip and normal faults, which appear on maps as more or less straight lines cutting the folds at various angles. Shortly before an overthrust dies out, it frequently branches into a number of smaller overthrusts and then passes into plastic deformation.

Since the competence of rocks differs in various parts of the earth's crust, the positions of the shear planes in the same deformation may be different in different rocks. Because of these variations, the planes of the thrust faults may acquire wavy curves and change their dip each time they pass from one type of rock into another. Observations reveal that thrust faults are usually more gentle in incompetent rocks and steeper in competent ones. Undulating thrusts, however, may also be observed in series of uniform composition. This is due to the fact, as established by M. V. Gzovskiy and Ye. I. Chertkova (1953), that on the appearance of any fault in the earth's crust, including thrusts, there is a change in the stressed state near the rupture and the "force" lines (trajectories of the principal stresses) are curved in such a manner that the further growth of the fault is not in a straight line but is also curved. Inasmuch as a thrust, like other major faults, is "collected"

from numerous small "embryonic" faults, the final composite fault (Fig. 289) has a wavelike or scalloped shape.

Thrust faults develop as the result of local concentration of the plastic slip during the time when the plastic deformation still continues. In other words, thrust faults develop locally, within the area of the plastic deformation. Consequently, the orientation of a thrust fault plane may in many cases change during further plastic deformation. A thrust fault

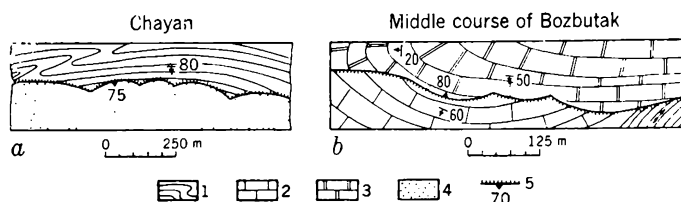


FIG. 289. Wavy shape of faults in nature (in the Kara-Tau region): (1) Carboniferous, (2) and (3) Devonian, (4) Ordovician (after M. V. Gzovskiy).

may become curved, for example, if the anticlinal axis changes its dip. Cases are known in which the thrust fault planes are folded; this indicates that folding continued after or during the development of the thrust fault. Figure 290, for example, shows folded thrust faults in the Kuznets basin. Such cases are rare, however, for thrust faults usually occur during the latest phases of folding.

It has been mentioned that strike-slip faults occur if the horizontal compression stretches the stratum horizontally. In addition to the small



FIG. 290. Folding of a thrust plane in the Kuznets basin (after A. Belitskiy).

strike-slip faults reviewed earlier, there are large strike-slip faults of regional extent that cut great parts of the earth's crust. These large strike-slip faults, like thrusts, are related to folding. Since they are due to horizontal tension along the strike of the fold, they most commonly have a nearly vertical dip and a strike diagonal to the folds. Strike-slip

faults usually occur along conjugate shear planes, and the combined total of the displacements leads to horizontal contraction normal to the folds and stretching parallel to them.

Like thrust faults, strike-slip faults die out horizontally and vertically, passing into plastic deformation, and they may have undulating fault planes. But their strike is more rectilinear than that of thrust faults. The displacements along large strike-slip faults may reach several kilometers, but most commonly they do not exceed several hundreds or even several tens of meters.

SHEAR

Let us consider faults that develop in the earth's crust as a result of shear. It should be said at the outset that shear in this case will be understood not as a structural form but as a mechanical process. It occurs as a result of the differential movement of one part of the earth's crust relative to another in any direction and deflects any given line in the zone between the two moving parts of the earth's crust. As extreme examples, we shall here consider the shear that causes either vertical or horizontal displacements along fractures that are nearly vertical in either case.

Vertical shearing in the earth's crust produces normal faults or reverse faults. The term *reverse fault* will be used for cases in which the hanging wall has moved upward relative to the foot wall. The mechanics of these faults are associated with a couple consisting of shearing forces, which consequently cause shearing and displace the adjacent blocks more or less parallel to the direction of their application.

Reverse faults are usually steep but not exactly vertical. The preceding chapter mentioned those faults formed in the marginal parts of a plastic body during its movement from below. We have seen that, in the case of the uplifting of a thick plastic body, the fractures first diverge and, during the continued elevation of the body, converge again, always remaining essentially vertical. Similarly developed fractures in nature are represented by the reverse fault planes that occur abundantly on the flanks of anticlinoria in mountain regions and the uplifted domes and arches on platforms. Reverse faults sometimes follow plastic deformation of the earth's crust. In the Tien Shan, for example (Fig. 68), anticlinoria form the basic structure; the reverse faults developed on the flanks elevate the stratum still further and complicate its structure.

Like normal faults, reverse faults are closely associated with flexures and pass into them wherever a reverse fault dies out. An excellent example of a reverse fault grading into a flexure, with a gradual decrease in the displacement, is to be seen in the Zhiguli Mountains: the middle part of their northern edge, where the displacement is greatest, is cut by a large reverse fault, and toward the ends of the Zhiguli folded zone the displacement gradually decreases; finally the reverse fault passes into a flexure. The diapir faults separating the elevated core from the surrounding rocks should also be called reverse faults. They usually have uneven profiles. As mentioned in the previous section, sometimes reverse faults that form the borders of the elevated cores of diapir domes are steep at depth and gradually become gentle, or even nearly horizontal, at the surface. If such a gradual decrease in the angle of dip of the fault planes also occurs on the opposite side of the ele-

vated core, the latter becomes mushroom-shaped or fan-shaped. This gradual decrease in the dip of the fault is due to the force of gravity and to the resistance of the overlying stratum. Under these conditions, the upper part of the block may slide horizontally to the side to form a mushroomlike shape. Earlier in this book, this was considered to be of great importance in the formation of general compression folds.

Reverse faults cannot always be distinguished with absolute certainty. If the faulting takes place near the boundary of essentially elevated or depressed regions, the faults cannot be called reverse because they are caused equally by the elevation and the depression. It is better to call them *normal faults*, keeping in mind the fact that the latter are not always caused by sinking but also by factors still undetermined. Vertical or nearly vertical faults are sometimes known in the course of oscillatory crustal movements to divide zones of intensive uplift from those of intensive subsidence, where these two are closely adjacent to each other. In these cases the relative vertical movement takes place along the fault plane, but direction of the movement may change from time to time. Such faults, with a long period of large and reversible displacements, have come to be known in the literature under the name of *deep faults* (Peyve, 1945). Their existence has been established in the Tien Shan, the Urals, and other regions. They determine the abrupt boundaries between major areas of the earth's crust, with sharply differing thicknesses of layers and different sedimentary facies. In particular, they are related to the process of oscillatory crustal movements. Deep faults of a different primary nature will be discussed later. Morphologically, deep faults may be treated as normal faults or upthrusts.

Shear deformation also occurs along the contacts of intrusive bodies during intrusion.

The significance of horizontal shearing in the earth's crust is still far from adequately understood. Large strike-slip faults are known in the Canadian shield (Joliffe, 1942). They extend horizontally for several hundred kilometers and also reach great depths. The so-called Talass-Fergana folded zone in central Asia is also considered to be an enormous strike-slip fault by many geologists. But the history of this folded zone is much more complicated; during the majority of tectonic phases the displacements were principally vertical, with relatively small horizontal shear. The great San Andreas fault zone in California, along which the earthquake of 1906 caused great destruction, is also called a strike-slip fault by many geologists. It is true that in 1906 there was horizontal displacement, but this event is only a small part of the fault zone's history, in which vertical displacements have prevailed and during the preceding long geologic period have produced a total displacement reaching several kilometers (Lawson, 1908). The data cited by many researchers to prove the existence of horizontal shear faults

in the earth's crust, with very great displacements, cannot be regarded as convincing.

On the other hand, the significance of horizontal shear in the earth's history is a question of great interest, inasmuch as on its solution basically depend our conceptions of the processes that take place at depth. Individual cases of shear in any direction are very abundant in the earth's crust. Actually, the mechanics of any case of faulting are those of shear, and the latter's peculiarities are evident everywhere. The additional or secondary deformations accompanying faults will be considered later.

PHENOMENA ASSOCIATED WITH FAULTING

Faults are accompanied by a number of typical phenomena reflected in various structural changes in the adjacent blocks. The study of these phenomena is of great interest: first, for clarification of the mechanics of faulting and, second, because they can be used in mapping with the specific purpose of determining the peculiarities of a particular fault.

Some of the general features of faults are as follows. Only very small displacements can occur along a single fracture, and even in this case the fracture splits into smaller fractures at the ends. Large faults never occur along a single fracture; they always take the form of a fault zone whose width may vary from fractions of a meter to hundred of meters. Such a zone consists of a large number of fractures with more or less parallel strikes, or of fractures intersecting and displacing each other at sharp angles, both vertically and horizontally. Owing to this conjugation of fractures, the fault zone contains long, thin lenses of relatively undisturbed rocks enclosed within brecciated zones. The individual fractures that compose the fault zone do not extend as far as the zone itself: they die out both along the strike and the dip and are replaced by other fractures usually situated en echelon. This structure of faults is formed by their repeated reopening and renewal of movement. The zone develops by spurts or thrusts, each new impulse using another fracture in the zone, so that the fractures intersect or displace each other. Since the displacements occur intermittently and the fault zones cut through rocks of various physical properties, the width of such zones changes frequently. The same zone may be very broad and filled in by gouge or breccia in one area, but within a short distance it may narrow considerably and cause no structural changes in the adjacent blocks.

Tension fractures may be open or closed. The open fractures never remain empty; they usually are filled by breccia or gouge derived from local rocks, or by new substances. The new materials may have various origins. They may be carried in solution by ground waters, which in descending deposit new minerals on the fracture walls and gradually

close the fissures. In other cases, the fractures may be filled with substances brought in by thermal waters from below. As the result of such hydrothermal deposition the cracks may be filled by veins of calcite, quartz, barite, ore minerals, etc. Finally, the fractures may be filled by molten magma, whose solidification produces dikes of various composition; there also are dikes of neptunic origin.

The majority of open spaces occur in normal faults produced under conditions of tension in the earth's crust, whereas thrust faults, reverse faults and strike-slip faults as a rule are closed or filled by gouge and breccia. This difference in the fault planes is sometimes very noticeable in underground workings such as tunnels and mine shafts. Cutting through a normal fault presents a certain danger: commonly there is a flow of water along the open cracks, which may lead to collapse of the walls or roofs of the workings. On the other hand, the closed fractures of thrust faults and strike-slip faults may frequently be crossed in underground workings without even being noticed by the miners.

Alterations in the surrounding rocks near the fault planes can be caused by the friction created during the displacements. The friction first of all affects the walls of the fault planes. The walls become more or less polished and display scratches, striations, grooves, and scars. If there are hard crystals in the fault plane, such as those of ore minerals, they scratch the walls during their movement. In other cases, friction along the fault plane produces so-called slickensides on the walls, which thus become shiny and smooth. The slickensides are created by thin fillings of newly deposited minerals such as chlorite, sericite, epidote, serpentine, hornblende, and other minerals of the so-called stress mineral group. Actually, it would be more accurate to call these *slip minerals*: they are formed along the slip surfaces as the result of the crushing and consequent recrystallization of the powdered material under high pressure. The scratches, grooves, and scars on the fault planes are frequently used to determine the relative direction of displacement. Naturally, the data on these may be of great help in determining the type of displacement. But great care must be taken in establishing the direction of the net displacement, if one uses slickensides and other such criteria.

It has been said that displacements along large faults occur not along a single plane but along many fractures. It can be added that the direction of displacement may change from one period of movement to another. Although the geologic relationships and detailed mapping permit us to establish the relative displacement along the dip, such displacement has not necessarily been in one direction only. On the contrary, most commonly each slip has its own direction, differing from that of the net overall displacement; the general displacement is nothing but the net result of numerous individual slips. Even this concept

is not an adequate reflection of the entire complexity of the problem. During each particular tectonic impulse, displacements may occur along different fractures of the fault zone in different directions. It has been said that a fault zone consists of a number of lens-shaped blocks. Each of these lenses may be displaced individually in a different direction. The lenses may be rotated, and some of them may move vertically while others are displaced nearly horizontally.

As an example, we may mention the great displacement that accompanied the earthquake of 1906 in California. Direct observations revealed that the earthquake was related to horizontal displacement along the large San Andreas fault zone. The horizontal displacement along the fault zone reached 7 m, but the corresponding vertical component was only a fraction of 1 m. Geologic mapping, on the other hand, has shown that the San Andreas fault is a normal fault with a great vertical displacement. The violent tectonic impulse of 1906 is therefore nothing but a small part of the history of the fault zone: this particular impulse produced a horizontal displacement, but the large number of previous similar impulses in the fault zone's history have for the most part caused vertical displacements (Lawson, 1908). The above example shows that the general direction of the overall displacement along a fault zone can be established only by detailed mapping and study of the structures in the area where the fault zone occurs. The use of slickensides and other criteria of slip is necessary, but the final conclusion must be based upon a thorough study of their occurrence and orientation on large surfaces. It must always be kept in mind that they may reflect only one of a large number of pulses or slips and not the net result of all the movements over a long period of time.

The structural changes in rocks adjacent to fault zones usually take the form of fracturing, cataclasis, and recrystallization of powdered rocks. The fracturing near fault planes, unlike the general fracturing reviewed earlier, is called *local fracturing*. This term emphasizes its special relationship to one fault or another. Each displacement takes place under conditions of intensive friction, which produces shear within the adjacent rocks. This determines the distribution and orientation of the special fractures accompanying faults. Earlier it was pointed out that both tension and compression occur during shear; both are diagonal to the direction of the couple in shear. Thus shear produces both tension and shear fractures. The tension fractures are turned in the opposite direction with regard to the shear. One of the two systems of shear fractures is parallel to the fault plane, and the other is perpendicular to it.

These two systems of shear fractures develop on both sides of the fault plane. A very typical feature is the featherlike system of tension fractures, which extend from the fault plane into the adjacent blocks.

One of the two systems of shear fractures, which is parallel to the main fault, prevails and is expressed in the form of short fractures en echelon. This system, developed on both sides of the fault plane, causes differential displacement in the same direction as the main fault. The second system of the shear fractures is usually much more weakly developed.

All these fractures form a zone of fracturing near the fault plane. The width of the zone varies greatly, depending on the amount of the displacement, the depth of its occurrence, and the physical properties of the rocks involved. Similarly developed tension and shear fractures frequently occur near intrusive contacts, for the igneous intrusion produces shearing in both the viscous magma and the enclosing rocks. In any event, the feather fractures, or (as they are sometimes called) *friction fractures*, frequently occur near intrusive contacts both within and beyond the igneous body (Fig. 41).

Y-shaped faults and the mechanism of their formation were discussed earlier. These frequently accompany large faults and form subsidiary faults.

The crushing, powdering, and other changes in the internal structure resulting from dislocational metamorphism lead to the formation of tectonites around faults. A brief review of the various types of tectonites was presented in Chap. 8.

Genetic Classification of Faults

Chapter 7 contains a morphologic classification of faults. The same faults may now be classified on the basis of their origin. Such a classification will be based on the character of the mechanical load to which the earth's crust is subjected (tension, compression, shear). It is not proper to base such a classification on the causes of the above load, since this would lead the discussion away from the origin of faults, as such, to far more general problems concerning causes of tectonic activity as a whole.

A complete classification of faults, including both genetic and morphologic characteristics, would require a table of all the morphologic varieties given in the earlier chapter. Such morphologic characteristics include those which refer to the shape of the fault (straight, curved, etc.), its position (vertical, inclined, longitudinal, normal, reversed, overturned, recumbent), and its relationship to the bedding (limited to one bed, limited to a few beds, intersecting the entire series, normal to the beds, diagonal), to the strikes and dips of the beds (parallel, cross, diagonal, conformable, unconformable, positive, negative), to the folded structure (parallel, cross, diagonal, radial, concentric), to the flow structures and shapes of the intrusives (parallel, cross, marginal), and to the topography of the area.

Table 18. The Genetic Classification of Faults

I. Tectonic ruptures				
Type of load	Usual dislocation prior to faulting	Type of rupture	Mechanical variety of rupture	Geologic variety of rupture
Tension	Stretching of hard beds between more plastic rocks; domes, antecises, anticlinoria	Fractures	Shear and tension fractures	Fractures: stretching, bending
			Tension	Stretching, normal faults
		Faults	Shearing	Strike-slip normal fault Normal strike-slip fault
Compression	Folds striking perpendicular to the direction of compression	Fractures	Tension	Tension fractures
			Shearing	Initial stages of faulting
		Faults	Shearing	Overthrust Strike-slip fault Strike-slip thrust fault Overthrust strike-slip fault
Shear (in the mechanical sense)	Vertical and horizontal flexures	Fractures	Tension	Featherlike or scar fractures
			Shearing	Initial stages of faulting
		Faults	Shearing	Upthrusts Strike-slip faults Upthrust strike-slip faults Strike-slip upthrusts

II. Nontectonic ruptures				
Process	Mechanical load	Type of rupture	Mechanical variety of rupture	Geologic variety of rupture
Contraction of rock's volume (due to cooling, crystallization, loss of water, etc.)	Tension	Fractures	Tension	General fractures
		Faults	Shearing	General fractures (rare)
Weathering and relief of load	Tension	Fractures	Tension	Fractures resulting from weathering and removal of overburden
Slides and avalanches	Tension	Fractures	Tension	Various fractures and faults
	Compression Shear	Faults	Shearing	
Extraneous processes (explosions, shocks, falling of meteorites, etc.)	Tension Compression Shear	Fractures	Tension	Various fractures and faults
		Faults	Shearing	

FAULTS OCCURRING AT THE PRESENT TIME

Faulting is also occurring at the present time. It may take place along already existing fault planes, or it may develop new ones. The displacements along fault planes rarely occur without violent disturbance, but one case is known in California in which the displacement takes place at an average rate of 6 cm a year along a low-angle thrust fault. The displacement first bends the oil pipelines, wherever they cross the plane of the thrust fault, and then breaks them.

The present faulting that takes place along existing fault planes or develops new ones in most cases causes earthquakes. One such case was studied during the earthquake of 1887 near the city of Verny (now Alma-Ata). The famous Russian geologist I. V. Mushketov, who studied the development of this earthquake, established that the focus of the earthquake was somewhere in the northern flank of the Zaalayskiy Range and along a normal fault that separates the mountain range from the adjacent plain north of the range. There were many avalanches along the fault plane (Mushketov, 1890).

Another violent earthquake in the same region occurred in 1911. Its epicenter was in the Kebin graben, which separates the Zailiyskiy Alatau range from the Kungey Alatau. This earthquake was studied by K. I. Bogdanovich et al. (1914). As in the preceding case, he established the relationship between the seismic impulses and fault zone (normal faults), showing that the earthquake produced a number of fractures with visible displacements. Some of these displacements created ledges up to 9 m high on the earth's surface. On the terraces of the Kebin River step faults were formed with displacements of several meters. Some of the fractures were several kilometers long and extended through both ranges and valleys.

The California earthquake of 1906 was studied similarly (Lawson, 1908). He established that the earthquake was associated with a violent displacement along the San Andreas fault, which trends through much of California for 900 km from southeast to northwest. A survey revealed that the earthquake was caused by a horizontal displacement along the fault zone; this displacement reached 7 m along the main fault and gradually decreased along the adjacent displacement planes. This was undoubtedly a discharge of the accumulated elastic loads. Comparison with the data from earlier surveys indicates that the stresses had been accumulating for at least the preceding hundred years, as a slow and unnoticed movement of the southwestern wall of the fault toward the northwest took place by means of elastic bending. Finally, the friction in the fault plane became insufficient to keep the adjoining walls of the fault plane together, and the sudden displacement occurred. The mechanics of this process can be easily understood from the following ex-

periment: two wax slabs are set down and pressed together. When moved horizontally in opposite directions, the slabs are deformed. For a while the plastic deformation continues without displacement along the plane in which the slabs are joined. Then the friction in this plane becomes inadequate, and the shear stresses are relieved by an instantaneous displacement of the slabs relative to each other. The horizontal slip along the plane of displacement produces a zone of tension fractures en echelon, clearly visible on both sides of the fault plane.

These and similar observations lead to the conclusion that earthquakes as a rule are the result of sudden movements in the earth's crust along fault planes. A number of other cases similar to those mentioned above have also revealed displacements visible on the earth's surface in the form of ledges, displaced roads, separated parts of buildings, and so forth. Table 19 (after Ye. F. Savarenskiy and D. P. Kirnos, 1949) is

Table 19. Amplitudes of the Visible Displacements during Some Violent Earthquakes outside the Soviet Union

Location of epicenter	Year	Length of active zone, km	Greatest displacement, m	
			Horizontal	Vertical
Wellington, New Zealand.....	1885	150		3
Sonora, Mexico.....	1887	60	...	9
Mino-Owari, Japan.....	1891	65-120	4	7
Beludjistan, India.....	1892	Less than 10	0.7	
Assam, India.....	1897	20	...	12
California.....	1906	450	7	
Island of Taiwan.....	1906	50	3	
Nevada.....	1915	30		5
Merigson, New Zealand...	1929	More than 4	...	5
California.....	1940	65	5	

a compilation of data on the size of the visible displacements along several fault planes with which a number of earthquakes outside the U.S.S.R. have been associated. The majority of earthquakes, however, are not accompanied by displacements on the surface. The displacements apparently occur somewhere at a depth of several kilometers, near the focus of the earthquake.

Improved seismic instruments and new methods of evaluating seismic data have permitted Soviet seismologists in recent years to begin a study of the mechanism of deep-seated earthquakes at their foci. This study reveals that as a rule earthquakes are connected with differential displacements of one part of the earth's crust relative to others and to shearing, in the mechanical sense of the word. Some earthquakes, especially deep ones, are associated with sudden compression within

the earth along a certain axis and elongation along another axis normal to the first. This deformation may be considered the result of shear along two systems of conjugate shear planes (Savarenskiy and Kirnos, 1949).

Study of various earthquakes has revealed widely varying directions of shear, compression, and tension in the foci of earthquakes. The great variety of the directions of displacement in the foci of the earthquakes corresponds to combinations of various types of displacements such as normal faults, strike-slip faults, thrust faults, and reverse faults. The same studies have revealed that the fracturing in the foci of earthquakes is not instantaneous throughout the entire area, but develops and spreads with a speed close to that of elastic waves.

Observations of the mechanics of earthquakes may provide extremely valuable data for interpreting the tectonic mechanism of faulting in general. On the other hand, everything known about the development of faults should be used in the study of the causes and conditions of earthquakes and the various phenomena accompanying them. The most active earthquake zones of the earth are located in belts of the most intensive oscillations of the earth's crust or belts in which movements of the greatest contrast adjoin each other. Such zones are known on the margins of the Alpine geosyncline, where the elevation of mountains and accompanying displacements in the earth's crust are still in progress. But some of these zones extend beyond the limits of the Alpine geosyncline into activated regions where intensive vertical movements in the earth's crust occur on platforms. Such regions are known in central Asia, southern Siberia, and East Africa. In Siberia and East Africa the areas of greatest seismic activity are located mostly in the large grabens, which are still sinking along the faults that border them (the Baykal and East African system of grabens).

The use of highly sensitive instruments has shown that along with the discernible earthquakes in tectonically active regions, there is also a large number of very weak seismic shocks. Scores and even hundreds of these may be registered in the course of one or two months over a comparatively small territory. If their foci can be determined with sufficient accuracy, the disposition of the latter can be used to establish the existence and location of tectonic faults that are hidden at great depth in the lower strata of the earth's crust and do not appear on the earth's surface. In tracing such deep faults one may also use more violent earthquakes, where these occur in large number, as, for example, in the zone around the Pacific Ocean. A similar attempt has also been made in the case of the southern shore of the Crimea, where a deep fault has been discovered that dips steeply from the Black Sea northward beneath the Crimean peninsula.

Seismologic investigations, which have become widespread in recent

years, have revealed interesting relationships between earthquakes, on the one hand, and the regional tectonic structure and history, on the other (Petrushevskiy, 1955; Belousov, Kirillova, and Sorskiy, 1952; Krestnikov, 1954; Gzovskiy, Krestnikov, Nersesov, and Reysner, 1958; Gzovskiy, 1957; Rezanov, 1958; Belousov and Gzovskiy, 1954; etc.). These relationships make it possible to use seismic activity as a certain secondary criterion in tectonic regionalization. This criterion may sometimes turn out to be extremely important. In certain cases, for instance, the disposition of zones of the epicenters and foci of earthquakes indicates a lack of conformity between the deep structure of the crust and the surface elements. In the Caucasus the positions of the foci mark the existence of deep north-south trends that are perpendicular to the present surface structure.

The American seismologist H. Benioff, analyzing the distribution of earthquakes of varying force in time, came to the conclusion that the present tectonic life of the earth's crust may be regarded as an alternation of periods in which elastic stresses are accumulated with moments in which they are released. In addition, it has been noted that the rate at which this energy is expended is maintained on the average for several decades and then changes sharply, and a new rate is maintained over a period of time. These transitional moments usually coincide in time with catastrophic earthquakes. Thus in some epochs a longer accumulation of stresses will be required to produce earthquakes of a given force than in others (Poldervaart, 1955).

Another interesting phenomenon is the "interlinking" of earthquakes. This is reflected in the fact that the temporary bursts of heightened seismic activity usually occur simultaneously in the most varying and far distant regions. There is no doubt that this opens completely new fields in the analysis of the present-day mechanical structure of the earth's crust.

CHAPTER 28

FAULTS AND TECTONIC STRESS FIELDS

The preceding chapters have shown that tectonic fractures are directly connected with the stresses in the earth's crust. Fracture joints arise normal to the axis of the greatest tensile stresses; shear joints arise along the surfaces of the greatest tangential stresses, allowing for a certain deviation from these surfaces owing to friction on the shear surfaces. Faults develop along already existing fractures, either through the force of gravity (settling along normal fault planes) or by the continuing action of the stresses. The distribution of the joints and faults, in relation to more general tectonic movements and the formation of more general structures, is determined by the fields of stress arising in the earth's crust. Such fields of stress may be called *tectonic stress fields*.

In view of the above, it is possible to make use of tectonic fractures in reconstructing the tectonic stress fields that existed in the crust of the earth at the time the given fractures were formed. Important additions to these conclusions can be derived from a study of folded deformations (Gzovskiy, 1954, 1956).

If morphologic criteria are sufficient to assign a given disjunctive dislocation to the category of tensile joints (open fractures with uneven edges that show no sign of displacement along the plane of the fracture), the position of the latter decisively determines the direction of the axis of maximal tangential stresses. If the tensile joints are parallel over a considerable area, it may be concluded that this area was subjected to tension. In this case, however, the tensile joints alone are not enough to determine the position of the axis of greatest compression, and hence of the intermediate axis. It can be stated only that they lie in the plane perpendicular to the axis of greatest tension.

If the tensile joints form an en echelon system in a zone that is oblique to the strike of the joints, this grouping is an indication of shear deformation produced by a couple. The directions of the forces in the couple are easy to reconstruct: they have acted "against the pile" in relation to the strikes of the tensile joints (Fig. 291). But one can

easily make an error in determining the plane in which the couple has acted if the observations are made only in a chance erosional section through the en echelon joints. For a correct interpretation, one must establish their position in three-dimensional space; if the joints are not isometric in their own plane, this fact must be taken into account in determining the disposition of the shear surface: the long axes of the joints are most probably parallel to this plane.

Shear fractures are usually closed and polished and have traces of slip. They occur in two interconnected systems. The line of their intersections is the intermediate stress axis, and the axes of greatest tensile and greatest compressive stresses are perpendicular to it and bisect the angles formed by the shear fractures. Ideally, the shear fractures are perpendicular to each other, and in the absence of any visible slip it is impossible to determine which is the axis of tension and which is the axis of compression. When there are traces of slip, these can be used to solve the problem: looking in the direction of the axis of greatest compression, one sees right shear to the right and left shear to the left; looking in the direction of the axis of tension, one sees left shear on the right and right shear on the left (Fig. 291). The solution is also made easier by the fact that under natural conditions the shear joints deviate some-

what from their theoretical position, in such a way that the axis of greatest compression bisects the acute angle between them and the axis of tension bisects the obtuse angle. In such a case, one's conclusion

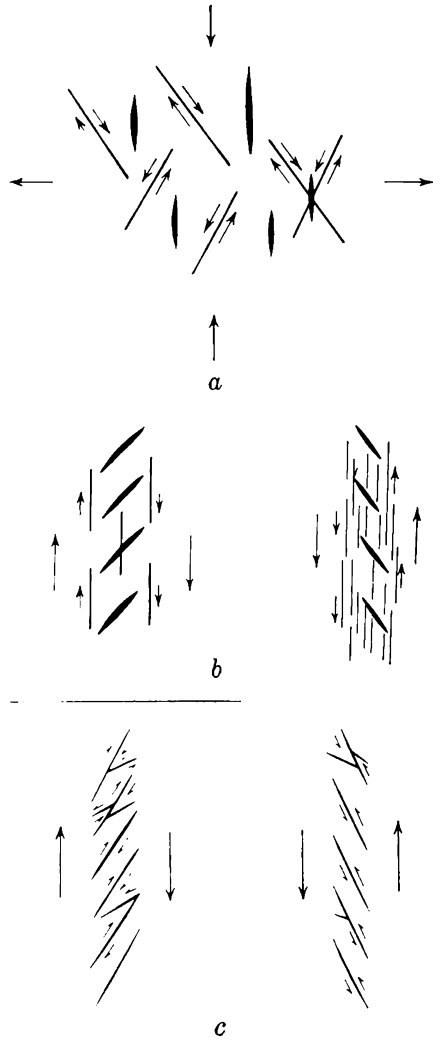


FIG. 291. The use of joints in reconstructing the tectonic stresses: (a) tension-compression; (b) shear deformation produced by a couple, with the formation of a system of en echelon joints; (c) the same, with the formation of a system of en echelon shear fractures. Heavy lines are tensile joints; fine lines are shear fractures; arrows show the direction of displacement.

should, of course, be verified from the direction of the slip along the fractures.

When tensile joints occur simultaneously with shearing joints, it is considerably easier to reconstruct the directions of all the stresses from the criteria described above. Theoretically, when there is a couple, the position of the shear joints is parallel and that of the tensile joints perpendicular to the plane in which the forces of the couple have acted. But because of the deviation of actual shear joints from their theoretical positions, there may be two systems of shear joints en echelon, one of which has its joints at a more obtuse angle to the direction of the forces in the couple than the tensile joints, and the other at a more acute angle (Fig. 291).

In using disjunctive dislocations to reconstruct the directions of the stresses and the nature of the deformation that has taken place, it is useful to distinguish groups of mechanically similar displacements subordinated to the general deformation as a whole. Figure 292 shows

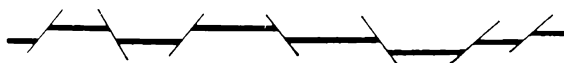


FIG. 292. Combined strike-slip faults.



FIG. 293. Combination of different kinds of faults. Black lines are compression faults; double lines are tension faults.

a map of strike-slip faults forming a single system, since all the strike-slip faults are associated in a single deformation: extension from left to right and compression from top to bottom.

Figure 293 shows a section through a combination of different faults that clearly belong to two systems: one system (solid lines) consists of fault displacements (thrusts) associated with horizontal compression; the other system (double lines) consists of fault displacements (normal faults) associated with extension in the same direction. It is clear that these two systems have been formed under different conditions of stress and thus at different times. In both these cases (Figs. 292 and 293), it has been assumed that we were dealing with simple displacements in the plane of the illustration. Actually, one must keep in mind the remarks made earlier in this book in regard to complex disjunctive dislocations, and in reconstructing the deformation from the fault displacements, one must not be satisfied with the impressions obtained merely from vertical and horizontal sections; one must establish the directions of the displacements more precisely.

The separation of the less plastic beds between more plastic ones into blocks, and the formation of such lenses, is an indication of extension. This is usually in the plane of the bed and normal to the straight boundaries, or planes of separation, between the blocks that have been displaced away from each other. In the case of cohesive formation of lenses, the lenses are turned at an acute angle to the direction of extension, as mentioned earlier. The lenses may be used to determine the relative amount of the tension in different directions. For this purpose, one must reconstruct the three-dimensional shape of the lenses, which are usually irregular ellipsoids. The short axes are the axes of greatest tension, and the long axes of the ellipsoids are the axes of least extension. But, again, one must stress the need to discover the shapes of the lenses in three dimensions instead of being satisfied with their outlines in a chance erosional section.

This discussion has concerned lenses formed by extension of a formerly complete, unfractured bed. Then in the direction of the greatest extension the frequency of the fractures (or the necks, in the case of lensing of the bed without complete rupture) will be greater than in the direction of the least extension; this fact also determines the respective dimensions of the lenses. This case also differs in principle from that of the flattening in compressed beds of any isolated objects (pebbles, fossils, etc.), which still remain whole and are not torn into pieces in the process of flattening. Here, too, the flattened bodies acquire irregular ellipsoidal shapes, but the long axes of these ellipsoids are in the direction of the greatest extension. Thus there may be two different orientations of the lenses, depending on the manner of their formation. If the lenses are the result of parting or separation (pulling apart) of a solid bed, they will be elongated in the direction of the least extension. In general compression folds associated with horizontal compression, with the greatest extension in the direction of the dip of the beds, the long axes of the lenses will be oriented along the strike of the folds. If the lenses have been formed by the flattening of small inclusions, their long axes will be in the direction of the greatest extension; that is, along the dip of the folds. Is this not the reason for the difference in the orientation of the crystals in rocks crumpled into folds, which is determined by petrologic methods and which has produced such a long and fruitless controversy?

In utilizing disjunctive and other dislocations, one must take account of the change in the three-dimensional position of the stress axes in different parts of the same structure. In the formation of a dome by vertical forces, for example, the axis of greatest extension will everywhere be parallel to the beds and the axis of greatest compression will be perpendicular to them. Consequently, the positions of these axes will change with any change in the attitude of the bed. Inasmuch as the

formation of general compression folds is accompanied by other deformations on various scales (bending and squeezing of the material in the flanks, injection and minor crumpling in the crests and troughs, general compression perpendicular to the axial plane, systems of small fault displacements, etc.), the distribution of stresses in the folds will be very complicated. The stress field along joints, faults, and any dislocations can be determined only for a certain part of the fold but not for the fold as a whole. In systems of deformations one can always distinguish the more general and the subordinate deformations, corresponding to more general and more particular stress fields. For example, a large intersecting thrust reflects stresses common to a whole group of folds; where the thrust is associated with a particular fold, it indicates the stress field on the scale of the fold; the breaking of beds into lenses

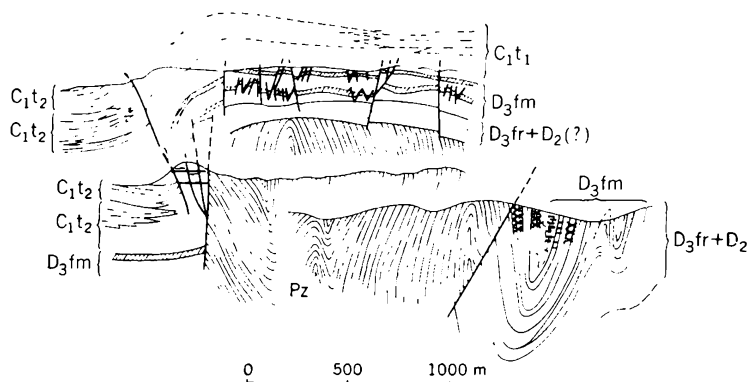


FIG. 294. Composite geologic section through a tectonic uplift in the Kara-Tau region (after M. V. Gzovskiy).

in the flank of the fold is determined by local stresses within the limits of this flank, etc.

It is extremely important to remember that stresses change in time as well as in space. They change with the development of a single tectonic deformation, even though the external forces may remain the same. In a geologic study of the structure of a part of the earth's crust in a mountain region, such as the Kara-Tau Range, which has undergone tectonic uplift over a great length of time, numerous systems of tectonic faults with very different orientations were found (Gzovskiy, 1954). In addition to these faults, which are visible in cross section (Fig. 294), there are also other faults formed later, which run across the trend of the uplift and are thus parallel to the plane of the illustration. These are very small normal faults. But the faults shown in the section can also be divided into several different types: along with the typical normal faults in the crest of the uplift, there are steep thrusts (reverse faults) in its flanks; the flanks also contain curious gently sloping thrusts with

the allochthonous flank displaced toward the axis of the uplift. This variety in the fault systems suggests that at different times this part of the earth's crust was acted upon by forces in different directions.

This structure was studied by various methods. All the faults were used in a direct reconstruction of the tectonic stress fields in different parts of the uplift. In addition, the photoelasticity method, which is widely used in engineering to study the stresses in machine parts, was employed. This method is based on the fact that certain elastic or optically active materials (such as gelatin) change their birefringence properties when the stresses within them are changed. A model of the geologic structure, made of such optically active material, was subjected to deformation similar to that observed in nature, and corresponding stresses appeared in it. Under polarized light in an apparatus similar to a petrographic microscope, by means of an optical compensator, or directly from the interference colors of an image of the model projected on a screen, it was possible to determine the distribution of tangential stresses of various relative magnitudes in the model (Gzovskiy, 1954, 1958). By this method, a picture of the distribution of stresses in a transparent optically active model of the uplift was obtained. Finally, the uplift was reproduced in a plastic clay model.

By all these methods it was established that a single gradual upward bending of the beds, without any other application of forces except those bending the beds upward from below, must produce all the observed systems of faults one after the other. Figure 295 shows the distribution of tangential stresses under these conditions and the directions of the greatest tangential stresses. The former indicates that the faults must have appeared first in the base and the crest of the uplift, whence they spread gradually upward and downward, respectively; the latter indicate the directions of the displacements to be expected in various parts of the uplift. The theoretical picture is clearly the same as the observed one.

Figure 296 shows the results of a reproduction of this uplift in a plastic model: the same systems of faults were obtained. The lateral faults, as explained earlier, are the result of rearrangement of the stress fields after the formation of the longitudinal faults.

The results of this study are of great importance in principle. They present the means of finding simple explanations for the formation of the numerous fault systems in the earth's crust, without recourse to any assumptions of changes in the directions of the tectonic forces. Without any change in the character of these forces, as the deformation develops there may be such changes in the stress fields as will produce considerable changes in the directions of the successively arising fractures.

In analyzing the formation of tectonic faults, one must also take into

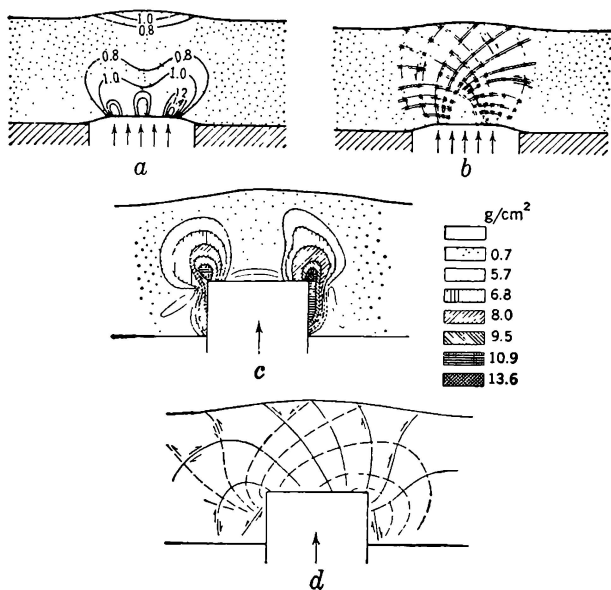


FIG. 295. Stresses and dislocations in models pressed from below by a die (after M. V. Gzovskiy): (a) relative magnitudes of tangential stresses in an elastic model, (b) orientation of shear surfaces in an elastic model, (c) distribution of tangential stresses in a plastic model, (d) orientation of shear surfaces in a plastic model.



FIG. 296. Plastic model of a tectonic uplift (clay); experiment by M. V. Gzovskiy.

account the history of such faults after their formation, during the course of the continuing deformation with which the faults were initially associated. If, for example, after the appearance of an echelon row of fracture joints, the action of the couple and the shear caused by it continue, the parts of the joints already formed will be rotated, whereas the newly formed parts of the same joints will be in the previous direction, corresponding to the stress field, which remains the same as before. This is the origin of S-shaped faults disposed en echelon, as depicted in Fig. 297. As the shear continues, the joints may be combined into a single great fracture whose convoluted, winding, or broken shape testifies that it was "put together" from many smaller fractures. There is also a deformation of the thrust surfaces in the further compression of the fold with which the given thrust is connected.

Later deformations of a different plan and not associated with the

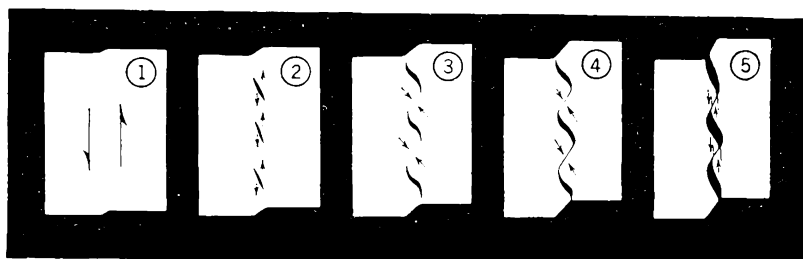


FIG. 297. Shear deformation (1) accompanied by the formation of en echelon fractures (2); the latter become S-shaped (3) and as the deformation continues are joined together (4, 5) (after M. V. Gzotsky).

deformation that directly caused the given fault are another matter. A tectonic fault, like any other surface in the earth's crust, may undergo deformation in the course of later dislocations. The bending of a thrust surface in the succeeding formation of folds has already been mentioned.

Another important factor, finally, in analyzing the mechanical conditions of the formation of tectonic faults, is the lack of uniformity of the earth's crust. Earlier, in connection with the rift valleys of East Africa, it was mentioned that faults sometimes deviate from their theoretical positions in following weaker surfaces within the earth's crust. Such weakened surfaces may be bedding planes, the contacts with intrusive bodies, and so forth. In their adaptation to these surfaces, faults sometimes diverge considerably from their expected positions.

The lack of uniformity of the earth's crust is partly the result of earlier faults. Cases are known in which fractures formed under certain mechanical conditions, such as shear joints, are used by fractures created later under different mechanical conditions, with a change in the direction of the forces, such as fracture joints. Thrusts produced in a stage

of compression may be reactivated in a later stage of extension as normal faults.

A. A. Sorskiy (1952) cited an interesting example of the superimposition of different deformations. He observed a bed that was broken up into lenses and crumpled into folds. Since the formation of lenses must have been due to compression normal to the bed and extension in its plane and the folding must have been due to the opposite disposition of stresses (compression along the plane of the bed), the entire deformation was divided into two stages occurring under different mechanical conditions (Fig. 298).

Thus the use of tectonic faults and other dislocations to reconstruct former tectonic stress fields must take account of many conditions and is a far from simple task. One method of overcoming these difficulties

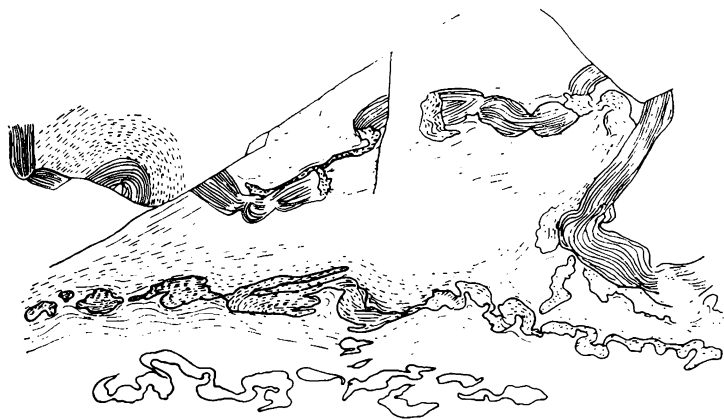


FIG. 298. Bed broken into lenses and compressed into folds in the Archean rocks of Karelia (after A. A. Sorskiy).

is to study tectonic fractures (usually joints) statistically. By a statistical treatment of the data on the occurrence of a great number of joints one can, of course, eliminate deviations due to local anomalies in the earth's crust or to minor succeeding deformations. But such a method is justified only in the case of joints associated with a single set of structural conditions. It can hardly give satisfactory results if it is applied indiscriminately to joints occurring both in the flanks of a fold and in all the other areas surrounding it.

The method of reproducing the distribution of stresses by photoelastic models was mentioned above. But a photoelastic model does not reproduce with sufficient accuracy the stresses throughout an object deformed plastically in nature. Thus the discovery of an optically active substance (ethyl cellulose), in which one may observe the distribution of stresses in the course of plastic deformation (Gzovski, 1958), is a considerable achievement.

CHAPTER 29

DEEP FRACTURES AND THE FUNDAMENTAL DIVISIBILITY OF THE EARTH'S CRUST

The concept of *deep fractures* in the earth's crust has been introduced into geotectonics (Peyve, 1945). These are very deep tectonic faults of great extent, which probably cut through the entire thickness of the earth's crust, and have a very long history. This means that the displacement along the deep faults, which are apparently vertical in every case, takes place over a long period of geologic time and is renewed repeatedly, so that the direction of the displacement may change. Along these faults some parts of the earth's crust may move upward at times and downward at other times, relative to the adjacent parts, for as long as a whole geologic period or even an era.

Thus movement along deep faults is one manifestation of the constant oscillatory movements of the earth's crust, but in this case the areas of uplift and subsidence are connected not by a flexure but by a fracture in the crust. As in other manifestations of oscillatory crustal movements, the regions divided by deep faults, depending on their movements (greater or less subsidence, or greater or less uplift), have different stratigraphic sections. Changes in thickness and facies associated with a deep fault are sharp and abrupt, so that on either side completely different geologic sections will be immediately adjacent. This close association between different thicknesses and facies in sedimentary series of the same age was explained earlier by the omission of the intervening sections as they were covered up in the formation of tectonic nappes, which in turn were believed to be the result of contraction of the earth's surface. In this older interpretation, many deep faults were regarded as the surfaces of enormous tectonic thrusts. The vertical position of the fault was thought to occur only at the earth's surface; at a certain depth it was supposed to become gently sloping or even horizontal. It has now been shown that these views are in error in most cases.

From the morphologic standpoint, the deep faults are vertical normal faults accompanied by fault zones and tectonic breccia.

It is usually possible to establish or assume a connection between the formation of deep faults and the contrast in the development of the oscillatory movements: where intensively rising and sinking parts of the earth's crust are very close to each other, a flexure is first formed at the boundary between them; this flexure is then replaced by a fault, along which the movements continue to occur. Such deep faults are *derivative*, or *secondary*.

Recently, however, the question of the existence of primary deep fractures, forming what may be called the *fundamental divisions of the*

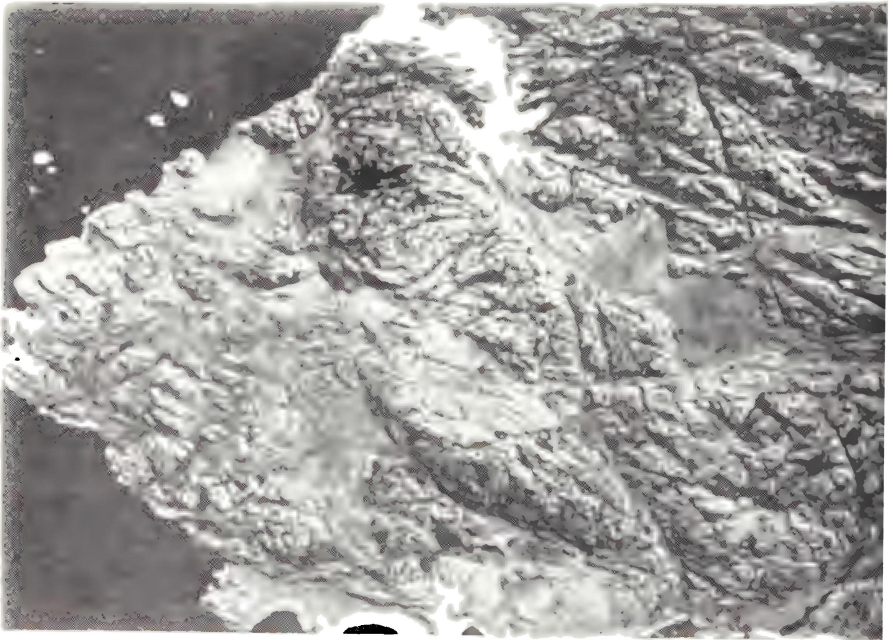


FIG. 299. Regional fractures in the Baltic shield, aerial photograph (after H. Cloos).

earth's crust, has been raised repeatedly. Figure 299 shows an aerial photograph of part of the Baltic shield, which is composed of crystalline rocks. It can be seen that the crust is here broken by numerous fractures in different directions. There is no possibility of connecting the disposition and intensity of this fracturing with any other more general movement of the crust, such as uplift of the Baltic region into a shield. Such fracturing of the earth's crust also exists in other places, and there is reason to think that it may be universal. It is manifested differently in different places, most clearly where the earth's surface is composed of crystalline rocks. It can also be seen in dense, consolidated sedimentary rocks. There is often a clear regularity in the disposition of these fractures. On the Colorado plateau of the U.S.A., for instance, over a vast

area of rocks with the greatest variety of structures, one can observe large faults trending north-south.

In some regions this fundamental divisibility of the crust appears indirectly, as revealed, for instance, in drainage nets. In some parts of the Hungarian lowland, which lies on a massif of metamorphosed and intensively dislocated Paleozoic and younger rocks covered by a thin mantle of sediments, the rivers follow surprisingly regular courses: they flow primarily north or south, and there are other areas in which they run east or west (Fig. 300). This disposition of the rivers may reflect an orthogonal system of regional faults. The regional faults in the crystalline basement are repeated in the sedimentary mantle, and they produce lines of weaker rocks that are later used by the rivers in their erosional activity.

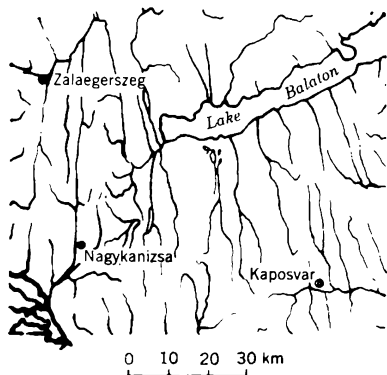


FIG. 300. River system in the vicinity of Lake Balaton in Hungary.

It has been suggested that fracture patterns on platforms, whose origin is unknown, are the reflection in the sedimentary mantle of the same regional fractures and that this is the reason for their regular arrangement. But the mechanism by which these regional fractures are "reflected" in the overlying sediments remains unexplained. The fundamental divisibility of the earth's crust is manifested chiefly in phenomena on a very much larger scale.

It has been noticed that in certain regions, over great areas, one can observe a predominant trend on the part of the major tectonic elements, both in geosynclines of different ages and on platforms. Such a predominance of definite trends is noted, for example, in a broad belt extending from Lake Baykal in the east all the way to France in the west. Within this belt, the major tectonic zones of subsidence and uplift and the boundaries between them follow a northwest-southeast trend, as seen in the eastern Sayan, the Altay, central Kazakhstan, the western spurs of the Tien Shan, the Kopet Dag, the Caucasus, the Donets basin,

the eastern Carpathians, the Dinaric Alps, the Apennines, and on the epi-Hercynian platform of Central and Western Europe, in which the synclises and anteklises are bordered by lines with a predominantly northwest trend. The same trend appears in Pay-Khoye and the Timan. It may be suspected that this extensive northwest trend of tectonic zones is the manifestation of some fundamental regularity.

A similar predominance of trends has been noted by many investigators (such as Staub, 1953). But it would be a mistake to try to extend a single system of tectonic trends over the entire earth's surface, since the predominance may differ in different parts of the earth. It should also be remembered that, in addition to regions where there is a distinct linearity or parallelism in the disposition of the major tectonic zones, there are no less widespread regions lacking such linearity, in which the tectonic structures have a tendency to form arcs or ovals. This is true of the western part of the Mediterranean Sea area, which is bounded by the ovals of the Alpine folded zones, and of the Carpathians surrounding the Hungarian basin and the mountain arcs of Asia Minor and Iran; the Himalayas, the Verkhoysk Range, and other chains of mountains also form such arcs.

In some cases one may suspect that the linearity of tectonic zones is historically a later phenomenon than their disposition in arcs or ovals. For example, up to the end of the Hercynian cycle (the end of the Paleozoic), the tectonic zones of Western Europe were disposed mainly in arcs (Fig. 212). The linear northwest trend of the zones outside the Alpine region in Europe appeared only after the beginning of the Mesozoic.

In central Kazakhstan, in the Caledonian cycle, the areas of uplift and subsidence had a predominantly north-south trend, but this predominance was not very great and, in general, the disposition of the zones in the Early Paleozoic was quite irregular. From the beginning of the Hercynian cycle, in the Devonian, the areas of subsidence (grabens) and uplift (horsts) had a northwest trend over the greater part of the territory. With the transition to platform conditions in the Alpine cycle, these trends were preserved in part, but to a considerable degree they were lost in the formation of the new "superimposed" round synclises (Fig. 301).

In central Asia the arclike curvature of the Pamirs (the Zaalayskiy and Darvakh Ranges) is combined with northwest and northeast trends in the western ends of the southern and northern Tien Shan; this is evidently associated with the basic divisions of the earth's crust.

The impression is thus created that the internal tectonic forces causing the uplift and subsidence of the earth's crust are adapted to its basic structural divisions, so that the orientation of the zones of subsidence and uplift follows these divisions. The impression is strengthened by

the intensification, in the vicinity of primary deep faults, of vertical crustal movements. One such case is the Talass-Fergana deep fault, which extends throughout the Tien Shan, from the Kara-Tau Range in the northwest to the Fergana Range in the southeast. Along much of its

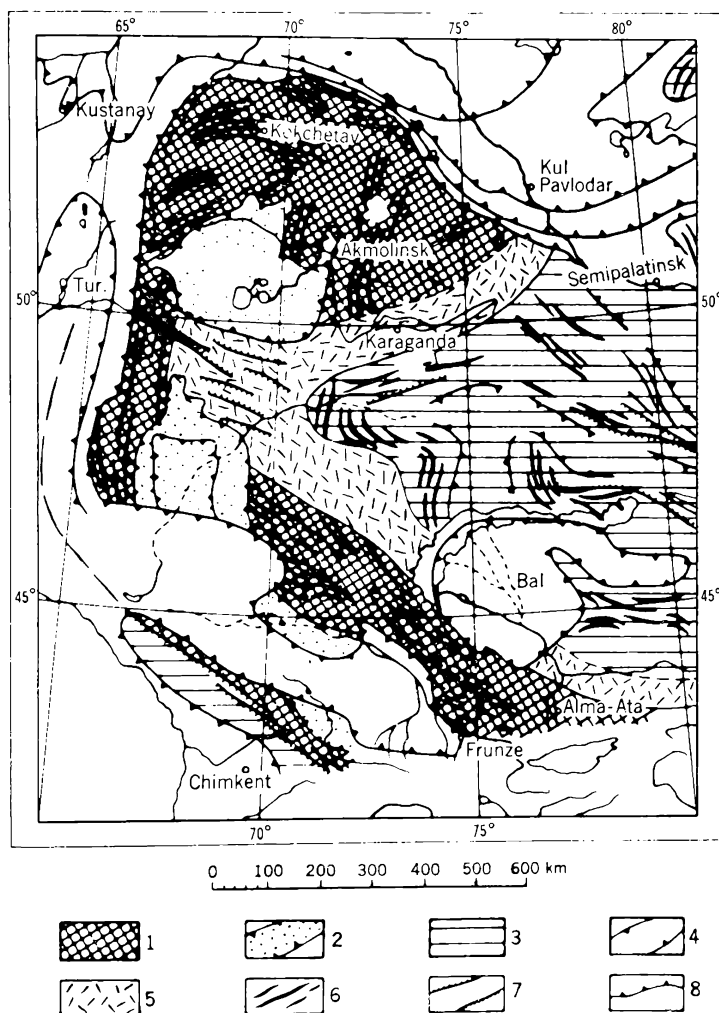


FIG. 301. Generalized Paleozoic structures of central Kazakhstan. (1) Caledonian zone; (2) Large depressions imposed on the Caledonian folded pediment and filled with Upper Paleozoic molasse; (3) Variscian zone; (4) Variscian synclinoria; (5) the marginal volcanic belt; (6) Caledonian anticlinoria; (7) major ruptures; (8) synclises and depressions of the Epi-Variscian platform.

length it separates the Caledonian geosyncline in the north from the Hercynian geosyncline south of the fault. In the region of the Fergana Range this fault, in bending southward, acquires a strike almost perpendicular to the general trend of the major tectonic zones in this part

of the Tien Shan, which is east-west. Along with this fault, another element almost perpendicular to the other tectonic elements of the Tien Shan is the Fergana Range, which closes the Fergana Valley in the east and is cut sharply on its eastern side by the fault (Fig. 302).

The history of the Fergana Range area shows that during the Jurassic period there was intensive and sharply asymmetrical subsidence; from the Fergana depression the extent of the subsidence gradually increased eastward as far as the fault, which bordered this zone of subsidence on the east. Beginning with the Cretaceous the region began to be uplifted, and the uplift again was asymmetrical: it increased gradually

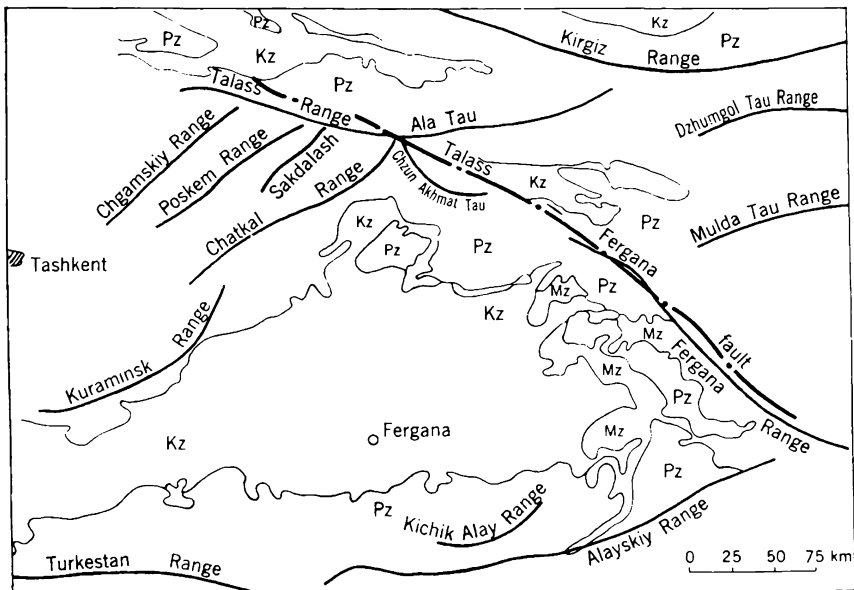


FIG. 302. Geologic and orographic sketch map of part of the Tien Shan.

from west to east, where there was a sharp boundary, again marked by a fault. Another interesting example is the structure of the Chelyabinsk brown-coal basin on the eastern slopes of the Urals; this is a deep asymmetrical zone of subsidence developing during the Triassic and Jurassic along a deep fault extending far to the north and south. These zones of subsidence and uplift next to faults seem to indicate that when forces deep within the earth attempt to draw the crust downward or to uplift it, they make use of primary deep faults as weaker surfaces along which to concentrate the most intensive uplift or subsidence, or both alternately.

The origin of these primary deep faults and of the primary divisibility of the earth's crust is completely unknown. It has been suggested that

these faults are associated with the stresses on a planetary scale that must arise in the earth's crust from the changes in the earth's shape, because of the gradual decrease in its rotational velocity. This entire problem, however, remains to be studied. In making further attempts to find a theoretical solution, it must be remembered that different trends of structures predominate in different parts of the earth's surface and that no one system of trends can be extended over all this surface. In the second place, a no less important circumstance is the appearance of a linear division of the crust in different places at different times. In the Archean Era there were apparently no linear trends, and the tectonic zones of subsidence and uplift had round outlines. It may be that the arcs and ovals of the tectonic zones in different regions in later epochs were the remnants of this Archean tendency. In post-Archean times, at different times and in different places, the rounded outlines of the tectonic zones were replaced by linear trends as the deep faults appeared.

The fundamental divisibility of the earth's crust and its blocklike structure are manifestations of the mechanical heterogeneity of the crust. In some cases the lack of uniformity in the crustal structure also appears in the formation of structural forms on a medium scale, such as domes, swells, and archlike uplifts. The fracturing that accompanies the uplift of these structures is not merely a function of the stress fields connected with the formation of the given structure but is adapted to the system of primary deep faults. Even the shape of the uplift may be determined by this net of primary faults.

The problem of primary deep faults is of essential importance for the principles of geotectonics, and it urgently requires study based on a thorough regional and historical analysis.

PART SEVEN

**IGNEOUS ACTIVITY AND
GEOTECTOGENESIS**

CHAPTER 30

INTRUSIVE IGNEOUS ACTIVITY

It has been shown that igneous activity is closely related to the tectonic history of the earth's crust, and a study of the history and nature of igneous bodies provides very important evidence for determining the processes at depth by which tectonic movements are fundamentally caused. Some aspects of igneous activity will therefore be discussed here, from the standpoint of their relationship to the structure and tectonic history of the earth's crust.

The relationship between igneous activity and geotectonics is still little known. For this reason, in many cases the discussion will have to be confined to the most general observations.

The manifestations of intrusive igneous activity have been classified as batholiths, fissure intrusions, sills, minor intrusives, and disseminated injections. The tectonic conditions of the formation and distribution of these intrusives will be considered below.

BATHOLITHS

The relationship between batholiths and their host rocks is of great interest. Batholiths that intrude into the earth's crust in great masses do not, as a rule, exert any considerable mechanical influence on the surrounding rocks. Figure 303 shows the relationship between a batho-

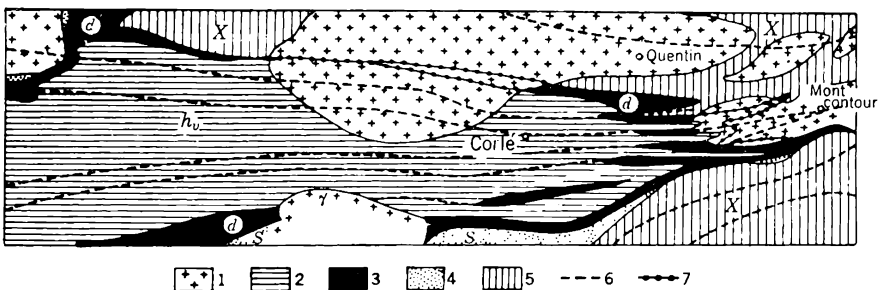


FIG. 303. Granitic massifs in the Corlé coal basin in France (after Haug): (1) granites; (2) Carboniferous; (3) Devonian; (4) Silurian; (5) Algonkian; (6) axes of anticlines; (7) position of Devonian deposits before the intrusion of the granites.

lith and the surrounding rocks in the Corlé basin in France. Here the sedimentary rocks at the batholith's contact are not bent or pushed apart: the batholith cuts through them, and they continue on the other side of the intrusive body.

Batholiths occur in zones where the beds around them are folded. These dislocations are not, however, related to the intrusive bodies, which intersect the beds but do not produce the dislocations. The relationship gives one the impression that a part of the surrounding rocks has been removed from the earth's crust without disturbing the structures of the surrounding stratum and that the void has been occupied by the intrusive body. There is no evidence of intrusion by force, as in the pushing apart, fracturing, or compression of the host rocks. This relationship between batholiths and their surrounding rocks was noted a long time ago, but the mechanism of a batholith's formation is still controversial. In general, there are two views. According to one of these, the ascending magma gradually makes room for itself by breaking off pieces from the stratum above it, which thereupon sink into the magma and are melted (Daly, 1933). This idea, however, is opposed by the observations of various authors who point out that the xenoliths of the host rocks in the batholiths retain their original attitudes and that their structures are related to those of the surrounding rocks. If the xenoliths had fallen into the magma, they would undoubtedly be encountered in various random attitudes. Moreover, some batholiths invade the surrounding rocks selectively: for instance, they penetrate farther along shale beds than along sandstones and limestones. This appears to be a result of the chemical action of the magma rather than simply of its mechanical action. It has also been pointed out that the difference between the specific gravities of the magma and the host rocks is too small to permit the latter to sink into magma. Thus another conception currently being elaborated by many authors is more likely to be correct. In this view, batholiths are formed less as a result of an invasion of the earth's crust by new magma than by melting and "granitization" of the sedimentary rocks in the locality of the intrusion.

E. Haug (1932) has suggested that the subsidence of the sedimentary strata at the bottom of a geosyncline brings them into a zone of high temperatures and active mineralization (by silica, water, alkalines, boric acid, and compounds of chlorine and fluorine). Through the exchange of molecules, the sedimentary rocks are transformed into granitic magma, which may be squeezed into the overlying strata by tectonic pressure.

At the present time, scientists are not inclined to consider high temperature as the principal factor in the formation of batholiths; it is thought rather that mineralizers penetrating the earth's crust under high pressures are far more important. The process of metasomatism

between these mobile and active solutions and the surrounding rocks gradually leads to the formation of granites in place of the previously existing sedimentary rocks. Thus the typically acidic and very uniform composition of the batholiths is therefore not their initial condition, but it results from a number of selective reactions between these volatile components of the magma and the surrounding rocks.

This discussion is not concerned with the petrologic problems of the physicochemical formation of batholiths, but it will note only that this process, whose existence is now accepted by many authorities in petrology, perfectly explains the observed relationships between batholiths and host rocks. The details of this process are described in numerous works on petrology. A review of this subject is presented in B. M. Kupletskiy's book (1942) and in the collection of papers entitled *The Problem of Granite Formation* (1949, 1950).

Naturally, a question arises as to the source of the energy that causes the volatile mineralizers to penetrate the earth's crust. This problem must be solved in conjunction with other problems of the energy in tectonic processes, and it will be considered later.

This view of the origin of batholiths assumes that the crust is extremely impermeable during their formation and can be penetrated from below, under high pressure, only by the most mobile substances. Such penetration under high pressure can be seen from the very intensive activity of the mineralizers and the great distance to which they have entered the earth's crust. These mechanical conditions of the formation of batholiths correspond fully to the known relationship between the intrusion of batholiths and other tectonic movements.

It is known that batholiths occur only among strata crumpled into continuous folds—that is, in zones of folding. They do not occur among flat-lying beds or in those which form only discontinuous folds. Even in zones of continuous folding, however, batholiths are not encountered everywhere. They are rarely found in the lower parts of the flanks of central uplifts, and even more seldom in areas of subsidence. Their typical location is the axes of central uplifts, where they form large masses within the preinversion complex of rocks.

This distribution of batholiths may be observed in any folded region. In the Urals they are elongated parallel to the eastern slopes of the range, along the margin of the Turgay interior depression, which apparently overlaps the central uplift. In the Tien Shan one finds large batholiths in the central uplifts of the Hercynian cycle, which coincide with the present ranges. An enormous batholith is associated with the Sierra Nevada central uplift in the United States, which arose in the Nevada intrageosyncline at the end of the Jurassic. Even greater batholiths are known along the Pacific coast of South America, where they are also located in the central uplift (the Andes). Similar intrusives

are abundant in the central uplifts of the Hercynian geosyncline in Europe: in the Moldanubian zone, the Spanish Meseta, etc.

Being associated with anticlinoria, batholiths most frequently occur along their most elevated parts—the tops of their crests—and disappear toward their plunges. Since batholiths are very closely related to the major positive structural elements, it is of extremely great interest to establish the exact time of their formation in relation to that of other tectonic movements.

Regarding the time of folding, intrusives can be divided into pre-folding, folding, late-folding, and postfolding categories, which differ in their internal structure and their relationship to the structure of the surrounding rocks.

Prefolding intrusives are completely formed and crystallized before all the phases of folding. They are folded along with the surrounding rocks and fractured, and they display signs of cataclasis and various secondary mechanical and physicochemical changes. Sills, and to less extent minor intrusives, form the majority of these. Postfolding intrusives most commonly fill the fractures in the earth's crust formed after folding, during the uplifting and fissuring of the enclosing rocks. These are fissure intrusives, to be discussed below. Their internal structure usually shows traces of movements of the magma, which are not, however, related to the dislocation of the surrounding rocks.

Batholiths, in this classification, are contemporaneous with folding and late-folding intrusives; in other words, they are formed while the folding is still in progress. Consequently, the fact that the folding is simultaneous with their intrusion and crystallization determines their internal structure and its relationship to the structure of the surrounding rocks. In intrusives that are contemporaneous with the folding, so that crystallization takes place at the same time as flow, the latter is fixed in the form of *flow lines* and gneissic texture. The varieties of orientation and their formation have been considered above.

Minerals become oriented because of the existence of a gradient of velocities: the individual streams of magma flow at different speeds and thus rotate the crystals and deform them. At the edges of intrusives formed simultaneously with the folding, the friction at the contact with the host rock creates greater velocity gradients and, consequently, a more distinct orientation.

The most abundant intrusives contemporaneous with folding are domical bodies in which the flow lines and orientation planes are parallel to the dome-shaped surface of the intrusives. This indicates that in the rising of the intrusive mass, the extension and flow were parallel to the domical bending of the surface, whereas the compression was normal to it. In other words, the mechanical conditions under which

these domical intrusives were formed are completely identical to those of the uplift of domes and extension of the beds in which they are formed. The flow structure within a gneissic dome therefore corresponds not only to the surface of the intrusive but also to the structure of the enclosing rocks. The overlying beds form a dome, with a periclinal attitude of the beds of the flanks.

It has been pointed out that intrusives frequently form not one but a number of domes of various sizes, as if they were superimposed on each other. This may be the result either of an irregular flow of the magma that might produce bulges and depressions in the surface or of repeated invasion of the same space by separate intrusions of magma.

Gneissic domes of this type must not be confused with laccoliths and other minor intrusives. In the latter cases, the magma, in being intruded, elevates and pushes apart the overlying rocks, whereas in intrusions contemporaneous with folding the flow of magma and the deformation of the surrounding rocks are both caused by a single and much greater general movement of the material in the earth's crust. The mode of occurrence of the gneissic domes is subordinated to the general structure of the area. Igneous and sedimentary rocks are frequently intermingled in the contact zone, where the scattered injections of magma create a zone of migmatization.

The gneissic intrusive domes in Saxony are of Hercynian and those in Scandinavia of Caledonian age. Intrusives simultaneous with the folding of the Alpine cycle, that is, formed at the same time as the Alpine folding, are rare; instead, Precambrian granite-gneiss intrusives are extremely widespread in this folded zone. They appear to predominate, and they are also closely associated with a gradual transition into the very small disseminated injections of magma that will be discussed below. It is quite likely that the great abundance of such ancient intrusives, which reveal signs of flow and crystallization simultaneous with the deformation of the surrounding rocks, is related to the greater plasticity of the earth's crust in Precambrian times.

Late-folding intrusives are distinguished by their nonconformable intersecting contacts and by the absence of flow, or by very few signs of flow, confined to the edges. Almost the entire intrusive body appears to be uniformly crystallized without any recognizable orientation of the minerals. The majority of large post-Archean intrusives are batholiths of this type. They include batholiths in the Urals, the Tien Shan, the Transbaykal, the Appalachians, and other regions.

The lack of oriented flow indicates that crystallization took place after the material had ceased to move. There are, however, clear signs that the intrusion occurred simultaneously with the movements at the basis of the folding deformations. This is evidenced by the fact that the forms

and positions of late-folding intrusives are closely related to the folded structures, despite the fact that the folds themselves are cut by the intrusives.

Batholiths are usually elongated ovals whose long axes parallel the trend of the anticlinoria. In a number of cases they are inclined in the same direction as the inclined folds. Another important criterion is the relationship of these batholiths to the surrounding rocks. In the first place, the overwhelming majority of batholiths show no signs of cataclasis and of the deformation and fracturing that would be inevitable if they had solidified prior to the end of the folding. (This observation holds true, of course, only if no later folding took place in the area.) Secondly, when the folded complex is unconformably overlain by younger deposits, the batholiths that intruded them are also intersected by the surface of the unconformity, even if the stratigraphic break has been very short. In Kazakhstan and the Tien Shan, for instance, the granitic batholiths that cut through the intensively folded sedimentary sequences of the Caledonian cycle are themselves cut by an erosional surface and overlain by the Devonian sediments that began the Hercynian tectonic cycle. Similarly, certain Hercynian batholiths in the Cordilleras of North America are unconformably overlain by Early Mesozoic sediments that belong to the Alpine tectonic cycle.

Thus both the batholiths with conformable internal structures and indications of flow and the uniformly crystallized batholiths that show no signs of the latter must have been formed while the folding was still in progress. They differ only in the relative time of their solidification: that is, in regard to whether the crystallization took place during the movement of the magma as a plastic body or after this movement had ended.

It has been observed that intrusives contemporaneous with the folding, the structures of which are in harmony with the structure of the enclosing strata, are frequently cut by later discordant and uniformly crystallized intrusives, which still, however, belong to the same folding period. In the case of a repetition of the tectonic phases of folding, such pairs of intrusives contemporaneous with folding and late-folding intrusives may also be repeated several times.

The entire mechanism of the formation of batholiths can be understood only if it is kept in mind that both the formation of the magma itself, by active mineralizing agents penetrating the earth's crust, and the movement of magma once it is formed occur simultaneously with the intensive uplift of the earth's crust in the intrageosynclines (central uplifts); this uplift is the basic factor in the formation of folds.

After solidification, the intrusives are fractured. The disposition and orientation of the fissures in batholiths and the conditions of their fracturing have been discussed in Chaps. 7 and 26.

FISSURE INTRUSIVES AND HYDROTHERMAL VEINS

This section will examine the formation of fissure intrusives along with the tectonic aspects of the formation of hydrothermal veins, inasmuch as both fill spaces of the same tectonic origin. The structural control in the disposition and time of formation of fissure intrusions and veins is governed mainly by the factors that determine the distribution of bending fractures. Such intrusives and veins are associated, as a rule, with convexities in the earth's crust: with anticlinoria, central uplifts, discontinuous folds and anteklises. Since fissure intrusives are always associated with large magma chambers, and the latter (batholiths and minor intrusives) occur principally within folded zones, the fissure intrusives and veins also occur within folded regions.

A large number of such intrusives, however, is known to be associated with large discontinuous uplifts invaded by intrusives (such as the discontinuous folds in the Rocky Mountains of North America). Fissure intrusives and veins are also formed on platforms, where they are encountered in connection with bending fractures developed in anteklises. This group includes, for example, the young Tertiary dikes and veins that cut the Precambrian crystalline rocks of the Canadian and Fennoscandian shields. The gold-bearing veins that intersect the Precambrian crystalline rocks of the Aldan anteklise on the Siberian platform are believed to be of Jurassic age.

The governing features in the disposition of the fractures relative to the shapes of the uplifts and the heterogeneity of their rocks have already been discussed. Veins frequently fill fractures along which displacement is evident. The products of igneous activity may completely penetrate the strata along faults (thrusts, strike-slip faults, normal, and reverse faults) bordered by a zone of abundant and intersecting fractures to form a complicated network of veins and veinlets. If these are particularly closely spaced, the entire fractured zone may become a single solid body such as a stock.

It is known that the fractures may become narrower or even die out. Consequently, the fissure intrusives and veins may also become narrower or die out, thus forming a chain of lenses of irregular shape.

Fissure intrusives and veins, along with the fractures they fill, develop after the folding in central uplifts, at the time of the intensive upwarping of the earth's crust in the epoch of mountain building. Consequently, the veins and fissure intrusives generally cut through beds crumpled into continuous folds and through the batholiths formed at the same time. This, of course, does not mean that they may not sometimes be found in bedding planes, cleavage surfaces, and surfaces of unconformities if these have become the sites of fractures. In other words, the heterogeneity of the earth's crust affects the distribution of such

intrusives and veins in the same manner as it affects the distribution of fractures.

As a general rule, fissure intrusives and veins are formed up to the time of the very large normal faults that also are associated with mountain building. This is explained by the fact that magmatic material may fill only those fractures that develop upward from below. Magma solidifies in such fissures at a time when they have not yet reached the surface and are thus blind. Normal faults, however, are the result of a further development of fractures and occur only if the fractures cut through a substantial part of the earth's crust or even reach the surface. Thus normal faults are usually dislocations dating after the mineralization,

whereas the veins under consideration here are formed at the same time as the ores (Spurr, 1916). There are, however, some exceptions in which veins appear to be associated with normal faults.

In attempting to determine whether a given vein is associated with a fault or a fracture, mistakes are often made by assuming that the cutting of one vein by another is the result of displacement of the one vein along the other, and thus looking for its continuation in the same direction, on the other side of the interacting vein (Fig. 304). Such a relationship between two veins may, however, be merely the result of the original position of the fractures. Fractures do not extend indefinitely in one direction but may be terminated by fractures running in some other direction. The question of whether

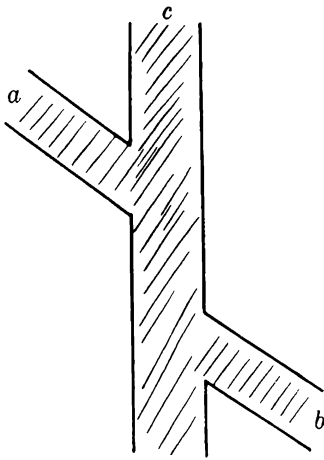


FIG. 304. Intersecting veins: (*a* and *b*) fractures of one system; there has been no displacement along *c*.

any tectonic displacement took place along a given fracture may best be answered from the attitudes of the sedimentary beds. If the answer must be based on the disposition of the veins alone, the analysis of the structure must be accompanied by a thorough study of the material composing the veins, their internal structure, and their relative ages.

The physical properties of the earth's crust during the formation of fissure intrusives and veins differ sharply from its properties during the formation of batholiths, when it was extremely impermeable. Later, as a result of its extension and fracturing, gaps are formed in it and its permeability increases substantially. Because of the great variety in the composition of the fissure intrusives, which range from extremely acidic to extremely basic, it must be concluded that during the formation of the fractures invaded by these intrusives there is intensive differentiation in

the magma chamber. The differentiation may be due to cooling of the magma or to a decrease in the pressure within the earth's crust resulting from its fracturing. Evidently at the beginning of the upwarping of the earth's crust, the material in the magma chamber is solidified near the edges but is still molten in its interior; from such chambers magma rises along the open fractures while differentiation continues in the chamber. It is sometimes suggested that the pressure of the magma itself causes the fissures in the earth's crust, or at least widens them. There is not sufficient basis for this view, however, since there is no evidence of any significant mechanical action on the rocks surrounding the fissure intrusives. An occasional bending or fracturing of the beds near such intrusions cannot be considered as such evidence, since the opening of fractures is the result of a much more general tectonic process, with which the intrusion of the magma is associated.

SILLS

Sills occupy a special position among the manifestations of igneous activity. They testify to physical conditions in the earth's crust sharply different from those existing in the formation of either batholiths or fissure intrusives. Magma does not move through the entire series of rocks, assimilating it as do batholiths, nor does it use the already existing fractures formed during the extension of the earth's crust; it invades along the bedding planes and fills them. Sills are typically of basic composition.

The great extension in area, along with the small and relatively uniform thicknesses of sills, indicates that the magma invades the bedding planes very freely. The earth's crust could not have been under any great pressure, since this would have prevented the flow of magma along the bedding planes. The intrusion undoubtedly occurs at relatively small depths. Moreover, the strata are easily separable and the magma enters along the surfaces between, pushing the beds apart as a knife would the pages of a book that are being cut open. Fissures extending far into the depths apparently serve as the feeders for these sills. Combinations of sills and fractures will be discussed below. It should be noted that magma may invade both bedding planes and interformational surfaces, as well as the surfaces of unconformities.

Sills have been observed between beds in both continuous and discontinuous folds as well as in flat-lying strata. The majority, however, occur in folded zones and predominantly in intrageosynclines. In the latter case, the sills, occurring between the folded sedimentary layers, are compressed into continuous folds along with the enclosing rocks. Thus there is no doubt that the sills were formed prior to the folding, while the subsidence of the earth's crust was still in progress. During

and for a while after their deposition, the beds are still unconsolidated, and the magma may easily be intruded between them.

Thus the sills in intrageosynclines are formed before the inversion, during the early stages of development of the geosyncline; consequently sills of this type are encountered only among the beds of the lower terrigenous sequence.

Such are the sills associated with the Early and Middle Jurassic deposits of the main Caucasus range. These sediments were deposited at the beginning of the Alpine geotectonic cycle, when subsidence of the earth's crust was still predominant, and the deposits were those of the lower terrigenous formation. The Devonian sediments with which the Hercynian cycle begins contain many sills in the Urals. In Western Europe sills occur in the folded zones of all three cycles. In the Caledonian cycle, as may be observed in the Caledonian structures of England, they are encountered in the Lower Paleozoic (Cambrian and Ordovician) strata. In the Hercynian cycle they appear among the Devonian sediments in the Hercynian structures of Europe. The sills of the Alpine cycle are contained in the Triassic and Jurassic deposits of the Alpine geosyncline. Many examples in other continents might also be mentioned.

In all such cases, as already pointed out, the sills are folded along with the surrounding rocks. Because of the folding and other processes connected with it, the sills are metamorphosed and turned into amphibolites or serpentinites.

Sills also occur on platforms; here they are associated with subgeosynclines. Some examples are the Devonian sills in the northern part of the eastern Russian subgeosyncline (in the area of the Ukhta River), the Late Paleozoic sills in the Tunguska basin, a large number of sills of Tertiary age in the Rocky Mountains, and the Triassic diabase Palisades sill near the City of New York.

MINOR INTRUSIVES

All the minor intrusives are hypabyssal bodies solidified at relatively small depths. The most common form is the laccolith. It is suggested that the bottom of each laccolith is connected with a feeder and the laccolith therefore has the shape of a mushroom. It has not always been possible, however, to see these feeders, so there is still some doubt whether each laccolith has such a channel. Some laccoliths appear to be completely isolated igneous bodies without connections to magma chambers.

At first glance, the formation of laccoliths seems, therefore, to be a puzzle. An explanation may, however, be found in the abundant and close relationship observed between laccoliths and sills. It can be

established in many cases that the laccoliths are local swellings of sills. In other cases, numerous sills extend in all directions as if they were very large apophyses of laccoliths: actually these laccoliths are the intersections of the sills. One example of the local swelling of a sill that thus forms a laccolith-shaped body has been described by M. V. Muratov (1945) in the northern Caucasus.

It may be supposed that true laccoliths are sometimes formed in this manner, so that they are thick sills of limited lateral extent or else the result of a local swelling of sills. The fact that laccoliths occasionally appear to be isolated bodies may be explained by later mechanical deformation of the sills in the same manner as the squeezing out of the fissure intrusives described above, which are also thus pressed into separate flat plates. In other cases, they may actually be mushroom-shaped.

Thus the mechanical conditions of the formation of laccoliths differ from those of the formation of the usual sills. Laccoliths occur when the earth's crust is not as freely penetrated along the bedding surfaces as in the formation of sills. Vertical compression in the earth's crust limits the injection of magma between the beds; and if such vertical pressure is not uniformly applied, it squeezes out the sills in certain places and allows them to swell in others, thereby forming laccoliths.

It is possible, however, that the majority of the intrusives called laccoliths are actually bodies of an entirely different type. In many cases, only the tops of the so-called laccoliths are accessible to observation. As a rule they are domical, or even composed of several superimposed domes similar to those of batholiths. The overlying rocks are partially elevated into domical shapes, but to a great degree they are broken through in the intrusion of the magma, accompanied by a number of mechanical manifestations at the contacts. Some such bodies pierce the overlying strata and become exposed to the air, not by erosion, but because they originally reached the surface in their formation (bysmaliths). The tops of these protruding bodies are broken into blocks, recalling the fracturing above diapirs.

It is difficult to establish exactly the full shape of these intrusives. It was mentioned earlier that V. N. Pavlinov (1946, 1947), on the basis of his studies of so-called laccoliths in various regions of the U.S.S.R., has concluded that these bodies actually are shaped like upside-down drops, or pears. Such "streamlined" forms of small hypabyssal intrusives, which recall the shapes of salt domes or diapirs, are evidently very widespread.

The formation of such drop-shaped stocks with domical laccoliths can be imagined only as the rising of one viscous substance within another, as in the formation of salt domes. Grout (1945) has described experiments to find an explanation for the shapes of certain intrusive bodies.

He allowed bubbles of a lighter fluid and of air to rise through a heavy fluid. During their upward movement the bubbles assumed one shape or another, depending on the relationship between the viscosities of the two fluids. They became elongated in relatively more viscous fluids and assumed streamlined shapes. It is quite possible that stocks, like these viscous bubbles, became completely separated from their sources and ascended through the plastic surrounding rocks as isolated "bubbles" of magma. Under these circumstances, their movement, of course, was extremely slow, and the magma had to overcome the great viscosity of the enclosing rocks. As the outer shell of the stock solidified, the movement was concentrated in the center and the magma continued to rise and finally pierced through the surface to form a bysmalith.

Minor intrusives are associated with a regime of intermediate tectonic movements. Consequently, they are encountered in transitional areas between platforms and folded zones and in the flanks of marginal and central uplifts and of foredeeps and intermontane basins. They also occur in some large discontinuous uplifts on platforms. Since these intrusives exert a mechanical action on the enclosing rocks, elevating them into domes, the combination of minor intrusives and intermediate or discontinuous folds is structurally reflected in the complication of the larger folds by smaller domes formed directly by the intrusives.

The minor intrusives are formed at the same time as the folds with which they are associated. Since the intermediate and discontinuous folds were formed over a long period of time, the formation of the minor intrusives cannot be fixed in time as precisely as the formation of batholiths, which are associated with relatively brief phases in continuous folding. For example, in the Rocky Mountains, the discontinuous folds were being elevated during the entire Tertiary period, so that one finds minor intrusives of Eocene, Oligocene, and Miocene ages.

Minor intrusives in the form of stocks of various shapes are also formed in intrageosynclines, during the stage in which the sills are intruded, before the inversion. Minor intrusives are predominantly of alkaline composition. They are formed in the crests of central uplifts at the very end of the geotectonic cycle, in the mountain-building epoch, simultaneously with the extrusive flows that terminate the cycle. At this time the geosynclinal regime has already come to a close and entered upon the transition to platform conditions.

Some examples are the minor intrusives of the northern Caucasus, which are trachyliparites of Late Tertiary age, usually called laccoliths, and which, according to V. N. Pavlinov, are drop-shaped stocks located in the Mineral'nyye Vody area, on the flanks of the Greater Caucasus anticlinorium where there are no continuous folds, on the margins of the foredeep. At the same time, these minor intrusives are associated with a transverse fold culmination that intersects the anticlinorium and ex-

tends northward, separating the Kuban and Terek structural basins. Minor intrusives of Mesozoic or Tertiary age ("laccoliths") in the Crimea also occur on the flank of an anticlinorium. Similar intrusives in the southern Urals and in the Kalba Range occur under the same structural conditions; these are of Late Paleozoic age, like the surrounding rocks. The Rocky Mountains are the classic region for the development of a large number of minor intrusives. This area lies between the folded zone of the Cordilleras and the North American platform. Such intrusives also occur in the marginal parts of the Caledonian folded region in Scandinavia. Near Oslo, for instance, they were formed during the Silurian and Devonian periods.

In the Kola Peninsula, near Khibinsk, they are associated with an antecline formed in the platform and are composed of alkaline peridotites and pyroxenites, nepheline-syenites, and ijolites. They are believed to be of Devonian age (Yeliseyev, 1937). Finally, a large number of minor intrusives of Mesozoic and Tertiary age are known on the platform of China, within a large area of intermediate folds of the same age.

The Late Paleozoic alkaline intrusives in the central Urals can be mentioned as an example of minor intrusives associated with the epoch of mountain building. Since the minor intrusives are formed during the uplift of the earth's crust, and themselves strongly uplift and extend the overlying beds, they are universally accompanied by fissure intrusives and veins.

A special group of minor intrusives of basic composition is associated with the flows of plateau basalts on platforms, both described as *trap-rock*. Such basic intrusives are common in the form of stocks on the Siberian platform, for example, among the Late Paleozoic flows of plateau basalts; they may be considered as "embryonic" flows. An intermediate link between intrusives and extrusives is the volcanic pipes filled with basic magma and volcanic breccia. These pipes contain the kimberlites—mica peridotites—famous for the diamonds found in them (in South Africa and Siberia).

DISSEMINATED INJECTIONS

Disseminated injections are probably the initial and deep-seated stage in the formation of batholiths, in which scattered injections of magmatic material gradually granitize the strata in the crust. This interpretation is supported by the fact that migmatization occurs abundantly in the immediate vicinity of batholiths, so that there is a transition from the disseminated injections of the migmatites, through larger intrusions, all the way to enormous batholiths. The scattered injections include granitoid rocks, pegmatites, and gneissoid granites.

It is probably incorrect to consider the zones of migmatization around

batholiths to be a result of the latter's action on the host rock, or as a network of their apophyses. On the contrary, it is most likely that the batholiths themselves were formed as a result of a complete merging of the disseminated injections and the assimilation of the rocks between them. The disseminated injections reveal the simultaneous occurrence of the intrusion of magma with the folding of the stratum much more clearly than batholiths.

Magmatic material invades the stratum along the bedding planes, schistosity, and other paths. At the same time, these injections show signs of plastic deformation undergone by the magmatic material along with the host rocks. During the folding, the injections of magma may flow from one place to another, become concentrated in certain parts of the stratum, and form lenses frequently bent into complex shapes or isoclinally folded. This may be observed both in microscopically small injections and in very large ones that may be transitional between the injections and batholiths. Thus there is a common flow of all the rocks, involving both the original stratum and the magmatic injections.

Scattered injections occur more rarely in rocks younger than Precambrian. In some cases, they may, however, accompany large batholiths in the central uplifts and in the preinversion complex of the interior depressions.

Depending on the extent of the absorption of the host rocks by the material of the disseminated injections and the shapes of these injections, one may distinguish bedding-plane migmatites, dendritic migmatites, migmatites in which the host rocks are contained as isolated blocks within the igneous substance, and "cloudy" or "shadowy" migmatites in which the pieces of host rock have been almost completely absorbed into the magma, so that they have lost their distinct outlines and can be distinguished only on the weathered surface of the work.

CHAPTER 31

EXTRUSIVE IGNEOUS ACTIVITY

Certain periods in the earth's history have been characterized by extensive igneous activity. Extrusive lava flows several kilometers thick and scores or hundreds of square kilometers in area are known, which were poured out within a relatively short time.

Geologic data indicate that effusions of magma to the surface, like intrusions, are much more extensive in geosynclines than on platforms. The history of extrusive activity may be divided into definite periods, which correspond fully to the periods in the history of the earth's movements. Each geotectonic cycle in the history of a geosyncline produces two series of extrusive rocks, formed at a different time in the cycle. The first is closely associated with subsidence of the earth's crust; it appears largely in the intrageosynclines and during the first, pre-inversion stage of geosynclinal development. The second series is associated, in time and space, with uplift. The very end of the tectonic cycle is the most typical time of the second series of effusions, and the central uplifts and stable intrageanticlines are the most common localities of their occurrence.

The first series of effusives, in the form of numerous lava flows, occurs in the lower terrigenous sequence—that is, in the same sediments with which the sills are associated. In chemical composition, the effusives and sills are very much alike, and they were undoubtedly formed of the same magma. The first series most commonly occurs as submarine flows. Like the sills, the flows contained in the lower terrigenous sequence are later folded with the entire sequence.

The intensity of the volcanic activity during the preinversion stage of the geosyncline may vary widely in different geosynclines. In many geosynclines, a thickening of the strata has been noted where the effusives occur. Since the underlying and overlying sediments usually have the same facies, the thickening indicates a deeper subsidence of the earth's crust in the areas concerned. This has been observed, for example, by L. S. Librovich in the Urals, by Richter in the mountains of the Celtiberian geosyncline, and by M. V. Gzovskiy and the present author in the Lesser Caucasus. In many intrageosynclines, however, the

intensity of the volcanism appears to be independent of the depth of subsidence. For example, the intrageosyncline of the Greater Caucasus sank deeper during the first half of the Alpine cycle but the associated effusives of the Jurassic and Cretaceous periods are relatively rare and small in volume, whereas in the Lesser Caucasus volcanic rocks compose almost complete sections of Mesozoic and Tertiary formations. Here intensive volcanic activity took place both before and after the inversion, during the entire cycle: before the inversion the effusions were concentrated in the intrageosynclines, to a larger extent in places where the subsidence of the earth's crust was deeper. After the inversion, volcanic activity first took place in the marginal part of the central uplift, then passed into the intermontane basins, and finally gradually died out. In the northern part of the Lesser Caucasus volcanism during the Jurassic period was concentrated in the Somkhet zone, which at that time was an intrageosyncline. In the Late Cretaceous, when the Somkhet zone was intensively uplifted, volcanism took place in the marginal basins formed north and south of the range. These are the southern Kura and Sevan basins; in the latter basin, the volcanic activity continued through the Eocene. The Tertiary and Quaternary volcanism took place first of all in the Araks basin—that is, in the central part of the newly formed intermontane basin.

There was extensive volcanism in the Ural-Siberian geosyncline during the first half of the Hercynian cycle. Here the Devonian and the Lower Carboniferous effusives made up the larger part of the preinversion formations in the Urals and especially in Kazakhstan, where the volcanism that began early in the Caledonian cycle (in the Cambrian and Lower Silurian) continued to take place.

Effusives are widely developed among the Lower Mesozoic rocks in the Alps, where they are associated with the Triassic and Jurassic formations and are called *ophiolites*. Consequently, the name *ophiolitic formation* has been accepted for a large number of formations that contain flows interbedded with the sedimentary strata.

There are effusives in the Lower Cambrian sediments in England, Scandinavia, and the northern Appalachians. They are everywhere associated with intrageosynclines, and with the latter's early stages. Nearly every geosyncline affirms the general rule that the intensive volcanic activity during the early stages of the geosyncline is related to its subsidence.

The chemical composition of the magma during the preinversion volcanic activity gradually changes: it begins with the eruption of basic or intermediate lavas and ends with acidic magmas shortly before the inversion. The latest phases of eruption reveal a rapid differentiation, as a result of which the composition of different flows differs widely, although acidic products prevail. For example, in the Urals the effusives

of the early phases of the Hercynian cycle (Lower Devonian) are chiefly composed of augite and pyroxene porphyries and their tuffs, whereas keratophyres, quartz keratophyres, and their tuffs prevail later. The Lower Carboniferous effusives are here most commonly represented by albitophyres and keratophyres; the volcanic activity finally ended in the early Viséan age. It is known that the inversion in the Urals occurred shortly before the Middle Carboniferous. A similar evolution of the magma occurred during the Hercynian cycle in the Tien Shan and Kazakhstan. The volcanic activity in the Lesser Caucasus also ended with the formation of albitophyres and other acidic rocks, although it began with basic and intermediate lavas.

It has been pointed out that the second series of extrusives is associated with the later stages of geosynclinal development—with the period of mountain building. Unlike the first series, which apparently ascended along fissures, the second series appeared chiefly through the vents of volcanos of the central type. They occur on the axial parts of the central uplifts, where, however, they are rarely preserved. The lava flows occupy the interior depressions and their margins. The second series occurs after the folding, and the lava flows overlie the folded stratum occupying the central uplifts (thus lying above all the other rocks of the cycle) or else are encountered among the formations of the interior depressions.

Examples of effusions that took place in the core of the central uplift are the Quaternary lava flows in the Greater Caucasus, associated with such volcanos as Elbrus, Kazbek, and others, as well as the Quaternary effusives in the Cordilleras of North America, the Andes of South America, etc. All these examples belong to the series of effusives appearing at the end of the Alpine cycle.

Examples of effusives belonging to the period of volcanic activity associated with the margins of interior depressions are the volcanos of the Mediterranean Sea area and the various volcanic island arcs: the Kuril Islands, the Islands of Japan, the Philippines, the Sunda Isles, the South Sandwich Islands, the Antilles, and many others.

To the end of the Hercynian cycle belong the Permian effusives of the Hercynian structures in Western Europe, where they occur largely within the interior depressions of the Hercynian central uplift. The Lower Mesozoic (Triassic) effusives of the piedmont zone, paralleling the Atlantic shore of North America, may also be included in this group, since they occur in an interior depression within the Hercynian geosyncline of the Appalachians. The end of the Caledonian cycle in Norway and in the northern Appalachians was also marked by various effusives.

The effusives of the second series are usually acidic to intermediate in composition (rhyolite, dacite, andesite), but the volcanic activity ends

with basalt flows. The chronologic succession of effusives in the Tertiary volcanic cycle in Nevada is as follows: hornblende andesite, hornblende-mica andesite, dacite, rhyolite, pyroxene andesite, and basalt. In the Hungarian basin it is hornblende andesite, pyroxene andesite, mica-hornblende andesite, rhyolite, and basalt. In many cases, this series ends with an extrusion of alkaline lavas.

If the region where the effusives occur had remained a geosyncline, the latest phase of the first cycle would have become the first phase of the next cycle. This suggests that in such cases it is better to speak of a single geosynclinal cycle, which began at the end of the first cycle and continued through the next. This concept of volcanic activity in geosynclines promotes a better understanding of the mechanical conditions of their development.

It is known that fissures form in the earth's crust at the end of each cycle. Owing to the fissures, the earth's crust becomes permeable, and this condition leads to eruptions of magma. By the time the subsidence of the next cycle begins, the fissures formed at the end of the first cycle still remain open and permit the issue of magma during the preinversion stage of the second cycle. The fissures gradually become filled with solidified magma and covered with newly deposited sediments. This decreases the permeability of the crust, and the eruptive activity gradually comes to an end.

In addition to geosynclines, however, outpourings of lavas occur (in some places very intensively) on platforms. Here many large flows of plateau basalts of uniform composition are known, such as the Late Paleozoic flows of Siberia, the Upper Cretaceous flows in India, and flows in Greenland, South America, etc. Apart from basalts, however, there are various other effusive rocks on platforms, such as the Tertiary and Quaternary volcanic flows of the central plateau of France, in the Rhine Mountains, and in Bohemia. In the central plateau of France one encounters flows of rhyolites, trachytes, phonolites, andesite-basalts, and basalts—the entire complex of effusive rocks, varying from acidic to basic in chemical composition. In the Rhine Mountains, alkaline rocks such as nepheline and leucite basalts, trachytes, trachytebasalts, phonolites, tephrites, and others prevail. Alkaline rocks are also very common in Bohemia, as well as among the effusives in the area of the East African grabens, where they vary widely in composition. Here the fissuring of the earth's crust was accompanied by outpourings of large amounts of lava and the formation of great volcanos.

As in geosynclines, the volcanic activity on platforms is associated with two types of tectonic regimes: first, with the subsidence of the earth's crust in subgeosynclines and, second, with its uplift and the formation of broad discontinuous folds. In the first case, the majority of the effusives are basic and emerge as flows through the fractures in

the earth's crust; in the second case, acidic intermediate and alkaline rocks make up the greater part of the effusives and appear in forms ranging in type from fissure flows to central volcanos. The first group includes, for example, the basalts of the Paraná subgeosyncline in South America and the Late Paleozoic effusive traprocks in the Tunguska basin. The second group includes the effusives of the central plateau of France that were mentioned above and those of the Rhine Mountains, Bohemia, East Africa, and the Deccan.

Mention must be made of the processes that take place in the foci of active volcanos. The ejection of large volumes of lava and gases from the depths of volcanos in some cases leads to the sinking of considerable parts of the earth's surface. Such is the origin of calderas of subsidence, which are sometimes tens of kilometers in diameter. From the purely structural standpoint, such calderas are circular grabens bordered by normal faults.

CHAPTER 32

GENERAL SURVEY OF IGNEOUS ACTIVITY. SOME GEOCHEMICAL PROBLEMS

GENERAL LAWS GOVERNING MAGMATIC PROCESSES

Let us attempt, on the basis of what has been said previously, to deduce some of the laws governing the development of magmatic phenomena and their relationship to tectonic movements. We shall first turn to geosynclines and trace the changes in magmatic phenomena in these during a geotectonic cycle. Here we shall begin not with the stage of subsidence within the geosyncline but with the period of mountain building that terminates the preceding cycle.

The great uplift that ends the geotectonic cycle in a geosyncline is accompanied, as we have seen, by intense discharges of magma on the surface. During the subsidence in the geosyncline (in the preinversion stage of its development), there are both flows and sills; at the end of this stage, small intrusions may be formed. Some time before the inversion, the igneous activity ceases. Later, during the period of inversion, the formation of the central uplift, and the main phase of folding, huge granite batholiths are intruded. Later still, from the beginning of the mountain building to the end of the cycle, fissure intrusives and ore veins are created and the whole cycle ends with a renewal of intensive outpouring of lava.

Examples supporting this scheme may be found everywhere. At the beginning of the mountain-building cycle in the Greater Caucasus—during the Lower and Middle Jurassic—many underwater extrusions and sills were formed. Many intrusives were formed during the Tertiary period. Finally, at the beginning of Quaternary time, there was another spate of intensive, but this time terrestrial, volcanic activity.

In the case of ancient cycles, there are often no visible traces of lava outpourings during the period of mountain building: this may be due to the removal of the products of terrestrial extrusion by erosion. But the first two phases of igneous activity—the initial extrusion and the intrusion associated with the inversion and uplift—always appear very clearly. They are well represented in the Caledonian and Hercynian

cycles in Tien Shan and central Kazakhstan, in the Hercynian cycle in the Urals, and in the Caledonian cycle in the Altay and western Sayan and elsewhere. The normal time of the submarine lava flows was the Cambrian and beginning of the Silurian in the Caledonian cycle, the Devonian and beginning of the Carboniferous in the Hercynian cycle, and the Triassic, Jurassic, and part of the Cretaceous in the Alpine cycle. The usual time for the appearance of granite intrusions and subsequent fissure intrusives was the second half of the Silurian and the beginning of the Devonian in the Caledonian cycle, the second half of the Carboniferous and the Permian in the Hercynian cycle, and in the Alpine cycle, depending on the time of the inversion, from the Late Jurassic to the Late Tertiary.

In this regular succession of magmatism, it is easy to see the connection between the type of igneous activity (extrusive or intrusive) and the state of the earth's crust, as determined by tectonic movements.

The period of mountain building is the time when the earth's crust is most penetrable, since at this period the crust is intensively fissured. This fractured condition of the earth's crust, as has been said above, is maintained to a greater or less degree into the succeeding period of subsidence within the geosyncline, so that the extrusions continue into this stage as well. But the deposition of new sedimentary strata plugs up the volcanic canals, which are also cemented by the magma that has congealed in them. This hinders the magma from emerging on the surface, and thus outpourings of lava are accompanied by intrusions. Since the sedimentary mantle has not yet been compacted and indurated and since the surfaces of separation between the beds are the most convenient paths for the intruding magma to follow, there is a formation of sills. By the end of the first cycle, the plugging of the fissures has completely shut off all extrusions. Thus there is a gradual decrease in the penetrability of the earth's crust in the preinversion stage of the geotectonic cycle.

The crust is least penetrable during the period of inversion. During this time, the crust in the geosyncline is subjected to strong upward pressure from below and is upwarped, forming a central uplift. This upwarping must, of course, overcome great resistance on the part of the generally highly rigid crust, so that the elevation is accompanied by considerable vertical compression and general solidification of the sediments. For this reason, the magma, though it is pushing upward, cannot penetrate either into the crust or through it. Only the most mobile volatile and active constituents of the magma are able to pass through the crust, as if through a sieve, and thus become the agents of granitization. These processes occur at the same time as the folding, which, as has been said above, is associated with the formation of the central uplift.

The folding finally comes to an end, and there begins to be an upward rising of the area of mountain building. The solidified crust of the earth becomes cracked, as the bending produces fissures. These do not follow the primary divisions between the beds but form new surfaces of separation in the rock strata, since after the solidification and folding the mechanical importance of the bedding surfaces is considerably diminished. Nonetheless, the lack of uniformity in the structure of the earth's crust influences the position of the fissures. Magma enters fissures that open up from below, forming fissure intrusives; these are accompanied by the formation of hydrothermal veins as well. Thus the fissure intrusives and veins correspond to the beginning of the period of mountain building. Later, as the uplift continues, the fissures become still greater and begin to cut through the earth's crust, so that fissure intrusions give way to extrusions. During this final stage, small intrusions of alkaline composition play a subordinate role.

This very simple outline may be more complicated in particular cases, mainly when the central uplift undergoes several phases of erosion. Thus batholiths, for example, may be created not all at once but in several stages. When there is a repeated renewal of intrusive activity, there may be a tendency toward a decrease in the basic composition of the intrusive with the course of time: the first intrusions will be of intermediate or sometimes even of basic composition, whereas true granite or granodiorite batholiths appear later. In the Urals, for example, according to B. M. Kupletskiy, two phases of intrusive activity are observed in connection with the Hercynian mountain building: biotite granodiorites were intruded in the first and muscovite granites in the second phase. In such cases, there is a superimposition of smaller magmatic cycles, which are often incomplete, on the greater cycle of igneous activity.

In addition, complications may be associated, as in the example of the Lesser Caucasus cited above, with the migration of igneous activity, especially a shift in the location of the lava flows. As a result of such migration, at the edges of the area (in the marginal basins) volcanic activity characteristic of the preinversion stage may continue while batholiths and fissure intrusives are already being formed in the central uplift.

It is especially interesting to note that the beginning of the uplift of the earth's crust within the intrageosyncline (the inversion) is simultaneous with the cessation of surface extrusions and that the further development of the uplift occurs at a time when the igneous activity takes the form of intrusions.

Shortly after the magma has broken through to the surface, during the period of mountain building, a new subsidence begins and continues

as long as there is extrusion of lava; this subsidence is generally more pronounced in places where a larger amount of volcanic rock has been extruded upon the surface.

One forms the impression that beneath the earth's crust, within the geosynclinal area, there is a constant accumulation of energy, which expands the melt beneath the crust and tends to push it out to the surface, and that this energy is expended as soon as the magmas have been able to emerge on the earth's surface. The loss of this energy of expansion through discharges of lava leads to a contraction of the material at depth; this contraction is manifested on the surface as subsidence of the earth's crust.

These observations will make it possible to understand why in particular cases extrusive igneous activity takes the place of uplift of the crust in the cycle. In actuality, the expansion of the mass of material at depth causes uplift and fracturing of the earth's crust, and finally, extrusion of lava. But one can easily see that in some areas the crust is more fractured, and thus more penetrable, than in other geosynclinal regions. Thus expansion at depth can immediately produce fracturing of the crust and discharges of lava without a great amount of uplift. The expenditure of energy from the depths is the same in both cases, but it is manifested differently. For example, in Alpine geosynclinal areas with an earlier (Mesozoic) inversion—in eastern and northeastern Asia and in the Cordillera—the uplift was, in general, less intensive than, say, in the Alps; but in these places, after the inversion and continuing to the Quaternary period, there were large outpourings of lava that did not take place in the Alps. In these cases, the Cretaceous and Tertiary extrusions took the place of the great uplift that occurred in the Alps; this substitution was evidently to a considerable degree determined by the fact that here the earth's crust was weaker and less able to withstand plastic deformation.

If we turn to the igneous activity on platforms, we shall find that it has many similarities to magmatism in geosynclinal areas. There are also differences, however, first in the scale of the phenomena and secondly in the smaller degree of regularity of the laws governing their development. It has been shown that subsidence of the geosynclines is sometimes accompanied both by extrusions on the surface and by the formation of sills. Thus there appears to be a certain succession in these processes, beginning under the conditions of a geosyncline and continuing under those of a platform. The volcanism, for example, that had begun in the interior basin during the time of large-scale oscillatory movements, continued for a while in the subgeosyncline that formed in the area of the basin.

It was further noted that discontinuous and intermediate folds are,

in many cases, accompanied by intrusions in the form of laccoliths and stocks, which later become fissure intrusives and veins. Fissure intrusives also accompany anteklises; as the latter are fractured and cracked, large extrusive blanket flows are formed over them. The similarity in principle of these processes with those observed in geosynclines is obvious. One of the most interesting differences is that, in regard to their position in the structure and its development, the batholiths in geosynclinal areas correspond to minor intrusions on the platform. Actually, since the batholiths are formed in the central uplift during the time of its erosion, the minor intrusions will accompany the erosion of discontinuous and intermediate folds, whose position analogous to that of the central platforms has already been pointed out.

In view of what has been said above regarding the circumstances under which laccoliths are formed, one must conclude that the mechanical conditions characterizing the earth's crust where this type of intrusive is formed are intermediate between those tending to the formation of batholiths and those producing sills. The penetrability of the crust is greater here than in the central uplifts. Because of this, not only the most volatile and mobile magmatic constituents but all the igneous melt as such succeeds in penetrating the overlying strata of rock. Under these conditions, the magma may even flow outward between the beds. But these conditions differ from those under which typical sills are formed: there is still a certain amount of compression, and the penetrability is limited.

It is worth recalling that the domical shape of many stocks is exactly the same as that of numerous salt diapirs and is in close harmony with the generally discontinuous nature of the structure. Here the magma, like the salt or gypsum, behaves as the most plastic element in the section and produces structural disharmony by its more active upward movement.

Various cases that at first glance appear completely similar may differ considerably in the intensity of the magmatic process. Although the full development of igneous activity is, according to this scheme, typical of geosynclines, areas are known in which they depart from the normal pattern. In the Welsh intrageosyncline, for example, and in the Celtiberian geosynclinal region, extrusive igneous activity is seen at the beginning of the cycle, but the batholith stage is lacking in both cases. The differences in the intensity of the preinversion volcanism in different intrageosynclines have already been mentioned. The reasons for these differences are not known. In any case, it would be a dangerous oversimplification to consider that the amount of igneous activity is directly or strictly related to the intensity of the tectonic movements. The relationship is doubtless much more complex, and at the present time not enough factual data are available to determine it.

SOME GEOCHEMICAL PROBLEMS

The last question touched upon here leads us into an area in which geotectonics and geochemistry merge and which, in spite of its great theoretical and practical interest, has scarcely been investigated yet.

It was said above that not only the forms of igneous activity but also the chemical composition of the magma change during the course of the geotectonic cycle. At the beginning of the cycle the extrusions, in general, evolve from more basic to more acidic lava. These extrusions at the beginning of the cycle together make up what is called the *ophiolitic*, or *keratophyrosplitic*, formations. These are followed by acidic, uniformly crystalline batholithic intrusions. Great variety has been observed in the chemical composition of the fissure intrusions; the texture of these rocks is usually porphyritic. The lava flows of the period of mountain building begin as intermediate and acidic and end as basic and alkaline. The magmatic cycle, which coincides with the tectonic cycle, is thus reflected not only in a cyclical succession of types of igneous activity but also in cyclical changes in the composition of the magma. All these magmatic manifestations in geosynclines, with the exception of those which end the cycle, are distinguished by the normal interrelationship between the alkaline and the alkaline-earth elements.

The less actively developed areas of the geosyncline (the intermontane basins and foredeeps and the parageosynclines), and also subgeosynclines of great extent, contain large amounts of alkaline rocks, which in the central uplift and interior basins appear only at the very end of the geotectonic cycle, when the geosyncline has already ceased to develop.

It has been known for a long time that the compositions of extrusive and intrusive rocks differ in different places and at various times. At first it was thought that these differences were purely geographic in nature—that is, that different geographic areas were characterized by the predominance of different magmas. This was the origin of the study of *petrographic provinces*. It was supposed that the earth contained two main provinces, the Atlantic and the Pacific, with a predominance of alkaline rocks in the first and of a “normal” sequence of rocks in the second. On the basis of these two principal provinces, various authors distinguished a large number of subsidiary petrographic provinces: within European Russia, for example, the Kola Peninsula, the White Sea, the Karelian, the Ukrainian, the Azov, the northern Caucasus, the Ural, and other provinces were delimited. Each of these is characterized by the predominance of igneous rocks of one composition or another.

Later, however, it was found that the geographic basis for the distribution of rocks of different compositions was only an apparent one. The chemical composition of magmas changes not only in space but also

in time and is closely associated with the tectonic development of the given part of the earth's crust. It turned out that rocks of the Pacific type were generally characteristic of folded regions and that Atlantic magmas were typical of strongly faulted regions on platforms or of areas that, having been folded earlier, were later subjected to uplift and fracturing.

This pattern governing the change in the chemistry of the rocks in association with the development of the earth's structure gradually became apparent. It is now known that alkaline magma is typical of platforms and of the transitional zones between platforms and geosynclines and that it appears in geosynclines proper only at the very end of the cycle. Calcalkaline magma accompanies all the typical stages in the history of a geosyncline. Since platforms are always formed on the sites of former geosynclines, observations in one locality will reveal the manner in which magma of "Pacific" composition changes with the course of time to magma of "Atlantic" composition. It is interesting to note that typical alkaline rocks are known only since the end of the Precambrian and that they occur in increasing amounts in the succeeding periods. This must be due to the fact that in the Precambrian, at least in the earlier part, the whole surface of the earth was characterized by geosynclinal conditions and that platform areas appeared only in the Paleozoic era (Levinson-Lessing, 1928, 1938, 1939; Afanas'yev, 1949; Daly, 1933; Goldschmidt, 1922; Niggli, 1923).

Observation of numerous examples has shown that basic magma in geosynclines, particularly in the form of extrusions but also as intrusions, is closely associated with and accompanies subsidence of the earth's crust. Acidic magma, on the other hand, primarily in intrusive but also in extrusive form, is associated in both space and time with uplifts. Moreover, it has often been observed that in the central uplifts of geosynclines, during the oscillatory movements of the earth's crust, acidic intrusions formed, whereas in the nearby subsided border areas basaltic lava continued to be discharged. V. I. Popov (1938) has pointed out some excellent examples of this in Tien Shan.

An additional illustration is provided by the outline of igneous activity in Europe given in Table 20 (Kossmat, 1936). This outline indicates that there are two granitic phases in each cycle, distinguished in this work as folding and postfolding. The porphyritic phase in the chart belongs partly to that of the fissure intrusives and partly to that of the extrusions of the period of mountain building.

It is interesting to compare Table 20 with Table 21, which was compiled by V. S. Koptev-Dvornikov to describe the history of igneous activity in central Kazakhstan. The periodic nature of the igneous activity is very clear in Table 20. Granitic intrusions, whose presence is

Table 20. *Cycles of Igneous Activity in Europe*

Phase of the cycle	Cycles of magmatism			
	Fennoscandian	Bohemian	Caledonian	Herzynian
Felspathic, alkaline rocks (after folding)	Alkaline plutons (Alne, Ragunda, Kola)	Lower Ordovician diabases	Alkaline basalts and phonolites in Car- boniferous of the Scottish graben	Triassic phonolites and alkaline basalts
Porphyritic, eruptive phase	Upper Algonkian por- phyries (Dala, Got- land), Rapakivi granite	Middle Cambrian porphyries	Porphyries and por- phyrites in Old Red Sandstone	Permian and Upper Carboniferous por- phyries and porphy- rites
Later granitic phase.	Archean granites (Stockholm, Nor- berg)	Algonkian granites	Silurian granodiorites, trondheimites, op- dalites	Carboniferous gran- ites, syenites, dior- ites
Earlier granitic phase	Laurentian granite- gneisses, leptytes, migmatites	Granite-gneisses and granulites	Deformed granite plutons	Tertiary tonalites and syenites
Ophiolitic phase.	Archean amphibolites	Middle Algonkian diabases and tuffs	Iotunian norites, peri- dotites, gabbros, dia- bases	Granite and tonalites in central gneisses (gabbros, peridotites, Middle and Upper Devonian diabases)

still unproved, have been noted in the Caledonian cycle. The fissure intrusives were omitted in all the cycles.

It has been said above that the lava flows at the beginning of the cycle are called ophiolitic formations. The word "formation" is often applied to igneous rocks. Igneous formations are associations of rocks that have a common origin, or, more accurately, a common connection with a definite stage of the tectonic cycle (as in the case of sedimentary

*Table 21. The Succession of Igneous Activity in Central Kazakhstan
(After V. S. Koptev-Dvornikov)*

Cycle	Time	Type of igneous activity	Composition of magma
Caledonian . . .	Cambrian	Extrusive	Diabases
	Tremadocian, Arenigian	Extrusive	Porphyrites, basalts, andesites
	Llandeilian	Extrusive	Dacites and trachyliparites
		Intrusive (sills)	Ultrabasic and basic rocks
	Caradocian	Extrusive	Andesites and acidic rocks
	Upper Silurian	Intrusive (batholiths)	Granites and granodiorites
	Lower Devonian	Extrusive	Basalts, trachytes, liparites
Hercynian . . .	Middle Devonian	Extrusive	Andesites
		Extrusive	Trachytes and liparites
		Intrusive	Granites (associated with part of the folding stage)
	Tournaisian	Extrusive	Andesites, dacites, liparites
	Viséan		Basalts, andesites, trachytes
	Middle Carboniferous	Intrusive	Granites, granodiorites, andesites, trachytes, liparites
	Permian-Carboniferous	Extrusive	Granites

formations). In this case, besides ophiolitic formations, one may also distinguish formations of granitic intrusions, vein or fissure formations, formations of alkaline minor intrusions, formations of Atlantic-type lava flows, etc. Various authors have described different formations on the basis of different principles of subdivision, which are often of purely local significance (Usov, 1945).

Although the final petrographic character of a given region is determined by the entire succession of igneous activity, later cycles of magmatic activity are reflected to a greater degree than earlier ones, since the vestiges of older cycles are less well preserved than those of later cycles. This rule is not absolute, however, and quite a few cases

are known in which the igneous rocks of older cycles are as prominent as those of younger cycles in the structure of the area.

The petrographic character of the locality also depends on the geotectonic element of the earth that predominated there during the last cycle and, to less degree, during the preceding cycle.

If the zone under consideration was an intrageosyncline during the last cycle, it will contain a normal sequence of rocks from keratophyrosplitic formations and granite batholiths to various vein rocks, etc. If another zone was a platform during the last cycle or cycles, the petrographic features will generally be determined, in the first place, by the rocks that were formed there during the existence of geosynclinal conditions, especially the last cycle of geosynclinal development; upon this background, in the second place, will be superimposed traces of igneous activity with the usual predominance of alkaline rock that are associated with the time when platform conditions characterized the given locality.

This is the picture in Central and Western Europe, for example. Here on the Alpine platform one finds, as already noted, similar areas of alkaline intrusive and extrusive rocks of fairly recent (Tertiary and Quaternary) age. But this part of the Alpine platform lies on a Hercynian folded basement, so that the alkaline rock cycle covers the older, still geosynclinal Hercynian cycle of igneous activity, which includes all the phases generally present in geosynclines, particularly that of large granite batholiths.

From what has been said one can see that drawing petrographic provinces without taking account of the ages of the igneous rocks (and their associations with given tectonic cycles), though it may serve certain limited purposes, is insufficient for scientific work and will inevitably lead to misunderstandings. Such provinces, of course, as that of the White Sea, where the igneous activity came to an end before the beginning of the Cambrian, should not be placed side by side with the Northern Caucasus, with its Late Tertiary igneous rocks.

If it were possible to establish the physical and chemical factors that determine the changes in the chemical composition of the magma during the course of a cycle, we would be far advanced on the road to understanding the causes of the deep-seated processes that lie behind tectonic phenomena. This task calls for the combined efforts of petrologists, geochemists, and structural geologists.

Metallogenic provinces are closely associated with petrographic provinces (Bilibin, 1948). This also presents complex problems to investigators. Different magmas, naturally, which correspond to different stages of the geotectonic cycle or to different structural zones of the earth, will result in the formation of different ores. The ores that accompany ophiolitic formations will, of course, differ from the ores formed in association with the granite batholiths of the period of inversion. Heavy

metals such as nickel, chrome, platinum, and iron are associated with the first; gold, copper, zinc, silver, lead, wolfram, tin, antimony, mercury, etc., with the second.

It appears, however, that rocks that are exactly or almost exactly the same in composition and correspond to the same stage of the cycle, but belong to different parts of a geosyncline of the same age, contain accumulations of different groups of ore minerals. These also occur in a sequence reflecting the connection of definite geochemical zones with definite structural zones. This circumstance is of very great practical importance and is used extensively at the present time. By establishing the connection of a given mineralization with a definite tectonic zone over a certain area and knowing the further extent of the same structural regime, one can, with considerable probability, also predict the distribution of the mineral resources.

The concepts of ore belts, zones, and regions have been formed in this manner. In many cases, it can be established that particular changes in the mineralization within a given belt are also controlled by different conditions of geotectonic development and by the different positions of the parts of the general structure.

From the standpoint of geochemistry, for example, the Caucasus is a particular region characterized by large amounts of such elements as barium, manganese, boron, arsenic, etc. At the same time the Caucasus is also distinguished as a separate structural province of the earth.

S. S. Smirnov has found two metallogenic belts in the Far East: an outer and an inner zone. The outer, western zone is characterized by Mesozoic folding and extensive granitic intrusives. Tin ores are typical of this belt. The inner zone, which touches the Pacific Ocean, is distinguished by Tertiary folding and by its small development of granitic intrusives but wide distribution of basic extrusives and injections. Copper ores are typical.

It is interesting that, within each of these zones, the changes in the complexes of mineral associations depend on the presence of anticlinoria or synclinoria, of complex or weak folding, of normal or thrust faulting, etc. V. I. Smirnov stresses the connection of ore belts with the tectonic structure of the locality and distinguishes belts that are (1) associated with faults, (2) situated along the boundaries between geologic zones of different mobilities, and (3) stretched out along supposed faults in the basements of folded regions (Smirnov, 1947).

There is no doubt that tectonic conditions affect the character and distribution of mineralization in various ways. In the first place, the mineralization depends on the lithologic composition of the host rock. But the lithologic composition of the rock, as is known, is determined by tectonic movements, especially oscillatory movements. In the second place, the mineralization is controlled by the structure of the rock. Ore

deposits are often connected with the arches of domes and of structural uplifts. This is due, on the one hand, to the fact that intrusions (both batholiths and smaller intrusions) are located primarily in structural uplifts. On the other hand, it is also known that the crests of uplifts produced by vertical forces are the most likely locations for the fractures that serve as channels for the passage of ore-bearing solutions to the surface.

In other cases, mineralization has also been observed to be connected with the fissures that accompany very large faults: thrusts, strike-slip faults, normal, or reverse faults. One example is the great fault in the Tien Shan, which runs through the Kara-Tau and Talasskiy Alatau all the way to the Fergana Range; this fault controls the location of a number of deposits.

In addition to these tectonic factors that control the positions of ore deposits, there is also the change during the course of the cycle in the composition of the magma, which has been described above and is not yet understood, as well as the fact that even the same magma in different places carries different ore-bearing solutions; these differences are connected with structural conditions.

From all that has been said, it is clear that our present knowledge of the processes that take place at depth below the crust and cause all the tectonic changes in the crust is limited. The further development of geotectonics and the extension of tectonic concepts to the depths of this planet depend on the progress achieved in studying the problems that have been touched upon in this chapter. Since these processes take place at very great depths, which are inaccessible to direct observation, petrologic and tectonic studies must be joined with geophysical investigations.

PART EIGHT

A GENERAL VIEW OF THE
GEOTECTONIC PROCESS AS A WHOLE
AND OF THE PRESENT TECTONIC
STRUCTURE OF THE EARTH

CHAPTER 33

TECTONICS OF THE OCEANS

GENERAL REMARKS

After examining all the various types of tectonic movements in the previous sections and determining the laws governing their development, we may attempt some generalizations that will provide a general view of the entire tectonic history of the earth and of the relationships between the various forms of tectogenesis.

One task remains to be accomplished, however, before this can be done. All the material that has been discussed above has been derived from observations made exclusively on the present-day continents. The ocean bottom and its structure and tectonic history have remained beyond the reach of our methods. Definitive conclusions, nevertheless, will depend on the light that can be thrown on these very problems. The oceans cover more than 70 per cent of the earth's surface, so that our conception of geotectonics may be completely different, depending on whether the laws of tectonic movements deduced from the evidence of the continents apply to the ocean basins as well or whether the oceans have developed entirely differently.

Some of the features of geologic processes in the oceans have been indicated in the section devoted to the oscillatory movements of the earth's crust. It has been said that the oscillatory movements in deep ocean basins, in contrast to the continents and shallow epicontinental seas, are not compensated by deposition and erosion. This distinction, however, refers only to phenomena on the surface and does not hold in regard to the development of the actual movements of the earth's crust and to igneous activity. Oscillatory movements, although not compensated by deposition and erosion, may still follow the same laws as on the continents.

An answer to this question is all the more necessary because there is at the present time no agreement among investigators in their attempts to solve the problem. Many geologists maintain the view that there is a complete separation between the geology of the continents and that of the ocean bottom; they assert that the oceans and the continents are primary structural forms of the earth and have totally different charac-

teristics and histories (see, for example, Schuchert, 1923). Other geologists hold the opposite view, that the structure and development of the earth's crust are, in principle, the same beneath the continents and the oceans (Kober, 1926; Stille, 1937; and others).

In recent years great successes have been achieved in efforts to solve this extremely difficult problem. The use of new geologic, bathymetric, and geophysical methods in numerous investigations in the Atlantic, the Pacific, and the Arctic Oceans has made it possible to learn some of the most fundamental features of the structure of the earth's crust beneath the ocean and to remove, if only partially, the curtain hiding the past history of the ocean bottoms. We shall attempt to summarize here the most important new information about the geology of the oceans and to draw some conclusions regarding their history and the conditions under which they were formed.

The following kinds of materials can be used to throw some light on the problem of the structure of the oceans: rocks brought up from the bottom of the ocean in the course of oceanographic investigations; the structure of oceanic islands and of those portions of the continents that lie next to the oceans; and geophysical studies, of which the most important are gravimetric and, especially, seismic. Deep seismic soundings made from ships enable us to determine the thickness and structure of the sediments on the bottom of the ocean, as well as the structure of the deeper parts of the earth's crust beneath the oceans. Study of the morphology of the ocean bottom is extremely important. Since the processes of sedimentary accumulation and erosion on the ocean bottom are comparatively slow-moving, and in the deepest parts of the ocean basins almost completely absent, the observed relief of the bottom is almost primary, determined for the most part by deep-seated tectonic and volcanic processes. Oscillatory movements of the earth's crust on the ocean bottoms leave direct traces in the form of uplifts and subsidences, the evidence of faults is retained in the form of benches, and volcanic processes are manifested in the volcanic cones that rise from the bottom. Thus the orographic method, which is almost useless on the continents because of the constant leveling of the topography through erosion and deposition, is not only appropriate here but plays a great role in forming our conclusions.

At the present time, the study of the structure of the ocean bottom has been carried furthest in the U.S.A. and the U.S.S.R. Most of the efforts have been devoted to the Atlantic and Pacific Oceans and the adjacent seas: the Caribbean Sea, the Sea of Okhotsk, and the Sea of Japan. Study of the Arctic Ocean has begun only recently, but it is being vigorously pursued. The Indian Ocean has been little studied thus far. The remarks that follow will pertain chiefly to the two oceans that have been most thoroughly studied.

SOME INFORMATION ON THE RELIEF OF THE OCEAN BOTTOM

The most prominent feature of the morphology of the bottom of the Atlantic Ocean is the Mid-Atlantic ridge, which extends from north to south along the center of the entire ocean (for the relief of the Atlantic Ocean bottom, see the literature: Bourcart, 1953; *A Discussion* . . . 1954; Heezen, Ewing, and Ericson, 1951; Kuenen, 1950; Tolstoy, 1951; etc.). Although the depths on either side are greater than 5 km, the depth of water above the Mid-Atlantic ridge is 2 km or less. In places individual peaks on the ridge rise above sea level, forming islands in a chain along the ridge.

On both sides of the Mid-Atlantic ridge are extensive deep basins with a generally level bottom at depths of about 5,300 to 5,800 m. East of the ridge are the Caspian, Angola, and African basins; west of it are the Argentinian, Brazilian, and North American basins. These are separated by fairly narrow ridges.

Detailed study of the relief of certain parts of the Mid-Atlantic ridge has shown that it is made up of numerous longitudinal ridges and depressions. In addition, the Mid-Atlantic ridge has been observed to contain great benches apparently produced by normal faults. There is an interesting ditch, or graben, that intersects the ridge laterally at about 30° north latitude; the depth of water in this graben is about 5.5 km.

On both sides of the Mid-Atlantic ridge there are broad zones of terraces, where the ocean bottom is divided into numerous level steps rising successively from the adjoining basins. These steps are as much as 50 km in breadth, and the height of one level above the next varies within the limits of several hundred meters. This terraced relief, which is probably due to displacements of the nature of normal faults, in places extends all the way to the deepest parts of the neighboring basins.

These very large, basic elements of the morphology of the ocean bottom are complicated by numerous submarine mountain peaks located for the most part around the large basins. Some of these mountains have the shape of sharp peaks, but many are huge truncated cones, with steep sides and cutoff, leveled, almost flat tops. Such truncated peaks, which are generally widely distributed over the ocean bottoms, have been given the name *guyots* (Hess, 1946). The flat tops of the guyots rise to various depths: 3 km, 1.5 to 1 km, 400 m. The tops of these table mounts are usually oval in outline and of various dimensions: some of them reach 50 km in length and 20 to 30 km in breadth. Almost everywhere that sufficiently detailed bathymetric observations have been made, the foot of each guyot has been seen to be surrounded by a ring-shaped trench several kilometers wide and hundreds of meters deep.

There can be no doubt that both the peaks and the guyots are extinct submarine volcanos. There is no other explanation for the origin of such isolated round conical seamounts rising to several kilometers in height. Moreover, the truncated tops of the submarine volcanic mountains can have formed only by ancient marine abrasion. Some of the evidence in support of this view will be cited below.

The greatest depths of the Atlantic Ocean have been found in the Antilles trench; here, near the island of Puerto Rico, a depth of 9,200 m has been measured.

The central part of the Pacific Ocean at the bottom is a vast flat plain at a depth of 5 to 6 km below sea level (see Betz and Hess, 1942; Bourcart, 1953; Chubb, 1934; Dana, 1875; Dietz and Menard, 1953; Gregory, 1930; Hess, 1946; Hess, 1948; Kuenen, 1950; Menard, 1953, 1955; Menard and Dietz, 1952; Murray, 1941, 1945; Stearns, 1945). This basic plain is broken up by a large number of subsidiary topographic forms, the chief of which is a number of steep terraces apparently produced by normal faulting.

Thus, for instance, the broad so-called Mendocino fracture zone extends westward from the California shore, separating the Gulf of Alaska in the north from the main part of the ocean in the south. This zone stretches outward along the 40th parallel of north latitude to a length of 2,000 km and a breadth of 3 km. Except for a small stretch near the shore, the southern flank of this zone is depressed along the entire length.

The Murray fracture zone extends in the same direction from southern California to the middle of the Hawaiian archipelago. This zone is 2,500 km long and up to 2,000 m wide in this case, the northern flank is lower.

The Clarion fracture zone extends outward from the continent to the west, at approximately 20°N for a distance of 4,800 km; in the opposite direction it cuts across the continent, where it is represented by a large fault and enters the Caribbean Sea. The southern flank is lower.

Still farther to the south, finally, the Clipperton fracture zone, the northern flank of which is depressed, stretches out in the same westward direction for about 5,000 km.

In addition to these fracture zones, the central Pacific plain is complicated by numerous island ridges and submarine heights. These ridges are linear, running mainly from the southeast to the northwest, although in some parts of the ocean (for example, from the Marshall to the Hawaiian Islands) there are submarine ridges that run in a perpendicular direction (southwest to northeast).

The islands of the central Pacific are usually of two types, volcanic and coral, although the bases of the coral atolls are also formed by submarine volcanic cones. The largest of the island groups—the

Hawaiian Islands—is situated on an enormous pedestal in the shape of a bank 1,900 km long and 600 km wide. This bank is bordered in part by an equally large depression in the relief of the ocean bottom. The main belt of coral atolls extends southeast-northwest from the Paumotu to the Marshall Islands. The atolls are grouped together on large bank-like upwarps of the ocean bottom; from these banks volcanic peaks rise to heights of several kilometers, and at the tops of the peaks are the coral structures.

Guyots—flat-topped submarine table mounts—are scattered widely over the Pacific Ocean. As in the Atlantic Ocean, these are usually oval-shaped, the longer axis of the flat tops being from 20 to 100 km and the shorter axis from 15 to 30 km long. The tops of the guyots lie at various depths below sea level: 400 to 500 m in the Gulf of Alaska; in the Aleutian depression a guyot has been found at a depth of 2,700 m, while at the edges of this same depression the tops of the guyots lie at depths of 800 to 1,000 m. Two guyots have been found west of California at depths of 600 and 800 m. In the vicinity of the Hawaiian Islands the depths of the guyots reach 2 km, although throughout the central part of the ocean their predominant depth is about 1,500 m. Around the Marshall Islands the guyots vary in depth from 1,100 to 2,200 m.

Guyots, as well as other submarine heights, have terraced slopes; such flat levels have been observed on the steep slopes of submarine peaks to a depth of 3,500 m. The greatest depths on the bottom of the Pacific Ocean, it is well known, are concentrated in narrow trenches that border the island arcs along the eastern edge of the Asian continent. On the slopes of these trenches, terraces have been found that were doubtless produced by faulting.

A subsidiary but extremely interesting feature of the topography of the ocean bottom is its submarine valleys (Bucher, 1940; Crowell, 1952; Dill and others, 1954; Ericson and others, 1952; Kuenen, 1950; Shepard, 1948; etc.). These submarine valleys are the direct continuations into the continental shelf of terrestrial river channels. They are the same as the latter in shape, size, and position, and there can be no doubt that they have been produced by terrestrial erosion. The fact that submarine valleys are very numerous in the continental shelves of every part of the world suggests that during some relatively brief and probably fairly recent period of time, the level of the water in the oceans stood lower than at present, so that the continental shelves were exposed. This was evidently during the Quaternary, when great amounts of water were tied up in the ice sheets and, according to calculations, the level of water in the world's oceans must have been 80 to 100 m lower than it is today.

Submarine canyons occur in the continental slopes in the deeper levels of the ocean, and even within the bed of the ocean itself. On the

continental slopes submarine canyons have been discovered almost everywhere that sufficiently detailed surveys have been made. They usually begin at the outer edge of the shelf and descend along the continental slope to depths as great as 4 km. On the bottom of the North American basin, submarine canyons have been found at the depth of 5.5 km.

The canyons associated with the continental slopes have extremely steep grades, as much as 50 m/km, which is much greater than the grades of terrestrial streams. Their longitudinal profile is complicated by steps that are even steeper. The canyon walls are usually precipitous and their bottoms relatively flat in cross section. The canyon walls are most often several hundreds of meters high, and the width of the canyons is several kilometers. The walls of canyons in the continental slopes have been found to contain rocks from the earliest geologic time to the Middle Pleistocene. The canyon bottoms are frequently paved with sandy sediments containing recent, often shallow-water, fauna. At their lowermost points the canyons terminate in broad sandy alluvial cones that have the appearance of deltas.

There has been a great deal of discussion of the problem of the origin of submarine canyons. All the facts known at present point to the conclusion that these topographic forms are not of terrestrial origin. It is impossible to imagine that during a short length of time (from the Middle Pleistocene to the present) the surface of the water in the ocean was lowered 4 km or more and then again returned to its previous level.

The longitudinal profile of the submarine canyons is another argument against their terrestrial origin: even during a short period of time (several hundred thousand years), terrestrial erosion of the unconsolidated rocks of the shelf and the continental slope would have had time to carve out profiles much more closely resembling those of terrestrial streams of water, whereas at present the profiles contain abrupt discontinuities at the edge of the continental shelf, from the underwater continuations of the river valleys to the submarine canyons. Finally, it appears that only a few of the great canyons are in such a location as to be the direct continuations of terrestrial rivers and their underwater channels. The majority of canyons are not connected with terrestrial rivers and are located in places where there are no great rivers emptying into the sea.

For these reasons, one must incline toward the view that connects the formation of the canyons with submarine erosion. It has been suggested that the erosion has been caused by "turbidity currents" running down the continental slopes—streams of water weighted by transported clay and sand particles. Such turbidity currents, with their high density, might have been created in large numbers during the ice age, when there was an abundance of melt water and when the continental

shelf was exposed, supplying the currents with loose material. The formation of turbidity currents would have been facilitated by the configuration of the shore line and the distribution of currents near the shore.

Although the submarine canyons were formed, apparently, at the time of the ice age, repeated studies have discovered changes taking place in them even at the present time; these changes consist mainly of avalanches and slides.

The hypothesis of turbidity currents explains the presence of interlayerings of sand that have been discovered by the latest observations among abyssal ocean silts, at depths to 6 km. In the Puerto Rican depression a layer of sand 2 m thick has been found at the depth of 8 km. Near Bermuda a sandy layer containing shallow-water shells was found among the most recent deposits, at the depth of 4,250 m. It must be concluded that such sandy interlayerings at uncharacteristic depths are detritus brought in by turbidity currents.

GEOLOGIC STRUCTURE OF THE OCEAN BASINS

Information on the geologic structure of the ocean bottom is still spotty and inadequate. Data on the geologic structure of the ocean basins have been obtained by direct study of specimens of sediment and rock brought up from the bottom, by photographing the bottom, by drilling near oceanic islands and continental shores, and by various geophysical investigations (see Carr and Kulp, 1953; *A Discussion* . . . 1954; Dobrin, Perkins, and Snively, 1949; Oliver and Drake, 1952; Richards, 1940; Shand, 1949; etc.).

According to seismic data, the thickness of loose sediments over the bottom of the open ocean is now several hundred meters, increasing considerably near the continents and islands. Up to 6 km of unconsolidated sediment have accumulated in the Puerto Rican depression. On the continental slopes the thickness of the sediments reaches 3 km.

Beneath the loose sediments are deposits that the seismologists have called *semiconsolidated*. These have the same distribution as the loose sediments, increasing in thickness near the continents and islands and decreasing toward the open ocean. Near Bermuda, for example, the thickness of the semiconsolidated deposits reaches 4.5 km, whereas away from the shore it falls to 500 m.

Geophysical and some drilling studies on the Atlantic shelf of the U.S.A. have shown that here the loose sediments, which are of Upper Cretaceous, Tertiary, and Quaternary age, increase in thickness as one moves from the dry land to the edge of the shelf from virtually nothing to 2,200 m. The semiconsolidated deposits (Triassic, Jurassic, and Lower Cretaceous) increase in thickness in the same direction from 0 to 4 km.

Thus the Paleozoic base upon which these Mesozoic and Cenozoic sediments lie must have a considerable downward slope toward the ocean. It must be noted that a large part of the Mesozoic and Cenozoic deposits of this area is made up of sediments of continental origin. Similarly, the basement of crystalline rocks that has been found on the other side of the Atlantic Ocean (on the Armorican massif in France, and in Cornwall in southwestern England) slopes downward toward the west in the direction of the ocean to a depth of 2,400 m at the outer edge of the shelf. The same downward slope of the Mesozoic and Cenozoic sediments toward the ocean has been found on the western shore of Africa, on the eastern shore of Greenland, and elsewhere.

In some places specimens of rock have been brought up from the bottom of the ocean. The specimens from the surface of the Mid-Atlantic ridge are almost exclusively basic extrusive and intrusive rock: basalts and gabbros. A specimen of limestone strongly altered to manganese was also brought up. The age of the basalt, as determined by the helium method, is Tertiary. Some of the specimens were tectonically worn rocks from fault breccia. The islands connected with this ridge are of volcanic origin and composed of basic and alkaline lavas; one of the islands was found to contain a large piece of granite enclosed within the lava. ~

South of the Azores, on the summit of a submarine mountain at a depth of 400 m, was found a limestone boulder with pteropod remains of Paleogene to recent age.

On a submarine bench near the Bermuda Islands, the present-day sediments at the depth of 1,500 m lie directly on calcareous Eocene clay. During the time between the deposition of this clay and the formation of the recent deposit, there must have been an interruption in the process of sedimentation.

Pebbles were found on one guyot in the Pacific Ocean, west of California, at a depth of 800 m. On another guyot, in the western part of the Pacific, were found the remains of Upper Cretaceous corals.

Some extremely interesting information was obtained by drilling holes in certain coral atolls in the Pacific Ocean. After a number of unsuccessful attempts, in 1952 on Eniwetok (in the Marshall Islands) two drill holes were sunk for the first time through the entire thickness of coral limestones into the underlying olivine basalts. These make up a volcanic cone, at the top of which the corals have erected their structure. Basalt was encountered at depths of 1,390 and 1,270 m. Study of the microscopic fauna has indicated that the coral limestones date from the Lower Eocene at the bottom to recent at the top; that is, the whole Tertiary system is represented. It appears, however, that Oligocene deposits are missing from the section.

Drilling on the Great Barrier reef of Australia shows that the coral

limestones are 150 m thick and lie upon the sands of a beach that existed here earlier.

At Bermuda the coral limestones extend to 75 m below sea level; beneath them are Miocene sands to the depth of 240 m, with volcanic rocks still farther down.

Thus an extremely important feature of the geologic structure of the ocean bottoms is the tremendous amount of volcanic activity. The entire ocean bottom is literally strewn with enormous volcanic cones, which at various times have evidently poured out huge volumes of lava. These extrusions are almost exclusively of basaltic composition. No volcanic activity of comparable scale is known on the continents. Recent data indicate that this intensive volcanism characterizes all the oceans, and not merely the Pacific, as was believed not long ago.

The origin of the volcanos and outpourings of lava is associated with clefts in the crust of the earth; consequently, the oceans appear to be areas of violent fracturing of the earth's crust. The presence of fissures is indicated by the linear distribution of many of the volcanic islands.

This linear distribution is especially noticeable in the Pacific Ocean. Here over a great area the islands and submarine volcanic peaks run in almost parallel ridges from southeast to northwest. Such island chains include the Hawaiian, Paumotu, Marshall, Gilbert, Ellice, and Cook Islands, the Fannin submarine ridge, and many others. Many of these islands appear on the surface as coral atolls, but it has already been indicated that the bases of these coral structures are volcanic cones.

Another feature of the geologic structure of the oceans is the continuation on the ocean bottom of certain continental structural elements of the earth's crust. The deep trenches that run parallel to the outer edges of island arcs (such as the Antilles, the South Sandwich Islands, the numerous island chains in the western part of the Pacific, and Indonesia) are, from the standpoint of tectonics, typical foredeeps, which, in contrast with those on the continents and in shallow seas, are not filled with sediments (Belousov, 1942). Cases have been noted in which the continental foredeeps are continued directly as ocean trenches (the Orinoco basin projecting into the Antilles trench, the Himalayan foredeep into the Java trench, and others). The suggestion has been made that the Mid-Atlantic and Mid-Indian Ocean ridges are very probably mountain structures formed in a young geosyncline of the Alpine tectonic stage; this is indicated not only by the relief of the bottom but also by the nature of the volcanic activity.

The discovery of the Lomonosov ridge in the Arctic Ocean is most interesting from this point of view. It is known that this stretches from the northern end of Greenland in a belt to the Novosibirsk Islands (10). The latter lie on the trend of the Verkhoyansk mountain chain, which

is one of the Alpine (Mesozoic) folded systems. At its other end, it appears, the Lomonosov ridge may be regarded as a continuation of the Mid-Atlantic ridge, which disappears north of Iceland in the shallow part of the ocean and at this point clearly bends westward, running partly under the Greenland ice sheet. It should be recalled that on the eastern shore of Greenland are deformed Cretaceous rocks.

It has also been noted that the flat basins at the bottom of the oceans, such as those in the Atlantic on either side of the Mid-Atlantic ridge, are evidently similar to synclises on the platforms of the adjoining continents of Africa and South America.

Thus it may be presumed that the major elements of the structure of the earth's crust, which are determined basically by its rhythmic oscillatory movements, are to some extent the same both on the continents and on the ocean bottom.

THE HISTORY OF THE OCEANS

It is well known that for a long time in geology there have been two competing views of the history of the continents and the oceans. Some consider that both have been constant features; others argue that they have changed during the course of geologic history. The adherents of the latter view, in particular, believe that in the past (in the Late Paleozoic and Early Mesozoic) there was a huge land mass in the southern hemisphere that was made up, in one combination or another, of the present-day continents of South America, Africa, Australia, Antarctica, and the peninsula of India. At the time that this land mass existed, it is thought, there was no ocean on the present site of the South Atlantic and Indian Oceans. Since it is also believed that until recent geologic time (to the end of the Tertiary period), dry land and shallow seas occupied the site of the North Atlantic Ocean, this conception is consistent with the idea that the Indian Ocean and the Atlantic are secondary oceans created fairly recently as a result of great subsidence of the earth's crust.

Only the Pacific Ocean is generally considered to be of ancient origin and to have existed throughout the entire course of known geologic history. This view, however, is also not unanimously held. Some investigators, among them E. Haug, believed that during the Paleozoic and Early Mesozoic there was a large land mass on the site of the present Pacific Ocean. The idea that continental land masses once existed where there are now oceans is based chiefly on analysis of the migrations of terrestrial fauna and flora, the most important of which is the *Glossopteris* flora of the supposed Gondwana land mass (see *A Discussion . . .*, 1954; Gregory, 1930; *Problem . . .*, 1952; Stearns, 1945; etc.).

In turning to this question, the first thing to note is the absence of any known oceanic sediments on the present-day continents. In spite of the fact that the surfaces of the continents have by now been quite thoroughly studied, no rocks have been found anywhere of which it can be said without question that they were formed in the abyssal zone of the ocean. Thus one may reliably conclude that, during the period included in the known history of the earth, there were no oceans on the present site of the continents.

Thus the arguments come down to the question: were there continents at any time on the site of the present oceans? There are numerous indications that in past geologic epochs there were large masses of dry land and shallow seas where there are now oceans. This evidence consists mainly of clastic material transported from areas that are now covered by oceans. Thus, for example, the distribution of Lower Paleozoic facies in the mountains of Scandinavia shows that, during the Cambrian and Ordovician periods, clastic material was carried into the sea, which at that time occupied the site of these mountains, from some highland to the northwest of Norway—that is, within the area of the present Atlantic Ocean.

It has been similarly established that during the Devonian period dry land existed to the east of the present continent of South America, again somewhere within the area of the present-day Atlantic. Thick continental deposits were accumulated in the Karroo basin in South Africa during the Late Paleozoic. A belt of these deposits, parallel to the Equator, is cut off to the west and east by the shores of the oceans; there is no doubt this originally extended for some distance into both the Indian and the Atlantic Oceans. In the Late Paleozoic and Early Mesozoic there was dry land south of the Caspian mountains, as indicated by the transportation of material from the south.

During the Late Paleozoic, Triassic, and Jurassic, a land mass was being eroded somewhere in the Atlantic Ocean west of the Congo basin. In the Late Paleozoic the land mass to the south, as is well known, was covered by a great ice sheet. From the evidence of the glacier's movement, it has been established that the ice in South Africa moved from the direction of the Indian Ocean, carrying boulders of Archean crystalline rocks. On the Falkland Islands the ice movement was also from the direction of the ocean—from the south. A similar occurrence has been observed in Australia, where the glacier descended from some dry land located somewhere to the southwest of Tasmania.

For a long time (the Paleozoic Era) large masses of dry land stood in the Atlantic Ocean east of the Appalachian system; during the same time there was land in the area of the Gulf of Mexico. During the Mesozoic there were highlands in the Pacific west of the Cordilleras and the Andes as well as on the site of the present Caribbean Sea. Until

the Tertiary period there was much more dry land in the area of the North Atlantic, and also in Indonesia, than at present.

In all the above cases, it has been impossible to determine the sizes of these areas of land that have since been covered by the waters of the ocean. But there can be no doubt that the continents were once greater than they are now; this is testified by paleontologic data on the migrations of fauna and flora from one continent to another. It is true that the most recent analysis of this problem compels one to doubt that a huge land mass, such as the proposed Gondwana land, could have existed at any time. But this same analysis shows that during the Paleozoic and Early Mesozoic the amount of dry land on the earth's surface was much greater than it is today and that there was a much better connection between continents, probably in the form of island chains and shallow seas.

The numerous indications of the recent deepening of the oceans and seas, which has already been mentioned, become especially interesting in the light of what has been said above. Thus, for instance, the drill holes on Eniwetok atoll show that the ocean in this area became deeper by 1,300 m during the Tertiary and Quaternary periods, since this is the thickness of the coral formations of these ages. There can be scarcely any doubt that the other coral islands of the Pacific have a similar structure and bear equal witness to the recent lowering of the ocean bottom. The broad belt of atolls from the Paumotu to the Caroline Islands almost certainly shows that this stretch of the ocean bottom has recently been greatly lowered relative to the level of the sea.

Since the flat tops of the guyots were, most probably, formed by abrasion, they are also evidence for the subsidence of the bottom and deepening of the ocean. In the center of the Pacific Ocean the guyots are now about fifteen hundred meters below the surface of the water; this is the same order of depth as the thickness of the coral structures. The deeper guyots are probably older and began to subside earlier. It is also very likely that the ocean bottom subsided unevenly: the deep guyots of the Aleutian depression are apparently the result of more rapid subsidence of the earth's crust in this area. If the terraces observed on the slopes of the guyots have also been produced by wave action, they are indications of even greater subsidences of the earth's crust maintained for longer periods of time.

Thus there is a basis for saying that the Pacific Ocean basin underwent great subsidence and deepening at a fairly recent period in geologic time. It is important to note that the level of the water in the ocean did not change greatly in relation to the adjoining continents, as shown by the identical course of repeated marine transgressions and regressions on the continents. It must be assumed that the lowering of the ocean bottom was compensated by the filling of the ocean basin with greater volumes of water.

There are signs of subsidence of the bottom and deepening of the ocean in the Atlantic as well: on the top of a guyot near Bermuda, as has been mentioned, Eocene shallow-water deposits occur at the depth of 1,500 m.

Attention has been drawn to the simultaneous downward inclination of the Paleozoic basement at the edges of the continents on either side toward the Atlantic Ocean. Since the Mesozoic and Cenozoic sediments that lie upon this base are either shallow-water or continental deposits, it follows that this inclination came about gradually and is a reflection of the same process of subsidence of the earth's crust beneath the oceans.

Indications of similar processes may be seen in many seas. The existence of dry land on the site of the Caribbean Sea during the Mesozoic was mentioned earlier; this land subsided during the Tertiary period (Eardley, 1954). Large areas of dry land existed during the Mesozoic and the Paleogene in the western part of what is now the Mediterranean Sea. From the transported clastic material it has been inferred quite reliably that this land lay to the west of the Apennine peninsula and north of Algeria and Tunisia; it subsided at the close of the Tertiary and beginning of the Quaternary periods. The large areas of land in the northern part of the Atlantic Ocean subsided at the same time. Very great and very recent subsidence of the earth's surface has been noted in the Indonesian archipelagos, where the formation of many interior seas and straits between islands took place almost before the eyes of man (Umbgrove, 1947). The same may be said of the interior seas of East Asia.

If all this is compared with what was said above regarding the existence of former land areas on the site of present-day oceans, the greater area of the continents, and the smaller depth of marine basins in previous times, a general tendency in the history of the oceans, despite the insufficiency of the data, becomes quite clear: expansion and, most important, deepening at the expense of the continents. There is reason to believe that the surface of the earth in earlier times contained no deep oceans and that their place was occupied by shallow marine basins smaller in area than the present oceans. With the passage of time the gradual subsidence of the earth's crust, occupying progressively greater areas and not being matched by deposition of sediments, led to the formation of the ocean basins.

This subsidence developed, apparently, mainly during the Mesozoic and Cenozoic; the clearest traces of it appeared during the Cenozoic Era in the Pacific Ocean. There this process is still actively taking place, possibly in association with the ring of intensive seismic activity that surrounds that ocean. In the Indian and especially in the Atlantic Ocean the same process in recent times developed much more quietly and now, perhaps, has ceased.

The subsidence of the ocean basins has been accompanied by fissuring

of the earth's crust and enormous basaltic extrusions. The cracking of the crust, as mentioned above, can be seen in the linear distribution of the volcanic cones in the central part of the Pacific Ocean.

The northwestern part of the Indian Ocean contains very clear indications of the collapse of great blocks of the earth's crust along fracture lines. It is well known that East Africa is the site of an enormous development of tectonic fractures in a continuous zone extending from the Dead Sea in the north to the mouth of the Zambezi River in the south. The character of the shores of Hindustan, Arabia, and the Somali peninsula leaves no room for doubt that they were produced by faulting. The high, precipitous, almost rectilinear eastern shore of Madagascar was also formed by a large normal fault. And there can be scarcely any doubt that the whole northwestern part of the Indian Ocean from India to Madagascar, with the extremely complex relief of its bottom, is a part of the earth's crust that has subsided and become an ocean bottom only in the most recent times.

SOME GENERAL PROBLEMS

The conclusion that there was a deepening of the ocean basins during comparatively recent geologic time compels us to raise a few general questions.

It cannot be thought that the increase in the volume of the oceans has been matched by the removal of an equal volume of water from some other source. The Cenozoic Era saw a general uplifting of the continents relative to the sea level; the displacement of a certain volume of water from epicontinental seas into the oceans caused by this uplift, however, is negligible in comparison and quite insufficient. The seas located on the present continents up to the beginning of the Cenozoic were very shallow, and the volume of water in them could in no way compensate for the deepening of the oceans. The same applies to certain local uplifts that took place simultaneously in the oceans, for example, in Indonesia, where certain islands were elevated 3,000 m during Quaternary time. Such islands are relatively tiny and displace negligible amounts of water.

There is no basis for assuming that subsidence of the ocean bottom in one place was compensated by equal submarine uplifts in others. In the Pacific, for instance, there is ubiquitous evidence of subsidence, but in the central part of the ocean, at least, there are almost no signs of uplifting of the bottom. The idea that the center of the Pacific was not subjected to seismic activity must also be banished: if there were intensive differential movements on the bottom of this ocean—a combination of great subsidence in some places along with great uplift in others—this contrasting movement would have to be reflected in seismic

activity. Thus it must be assumed that there was a general subsidence of the entire huge territory of the ocean as a whole, without its being divided into areas of different types of movement.

In the light of all that has been said, we must unavoidably conclude that there has been a great increase in the amount of water on the earth during the course of geologic time. Some investigators have recently come to this conclusion by other paths as well (Rubey, 1951).

The source of the water that filled up the ocean basins as they became deeper can be sought only in the depths of the planet. Water must have been given off by some transformations of the material in the depths. This view cannot be denied logically, since the main deep-seated process determining the history of the earth, as will be seen below, must evidently be the differentiation of matter, producing a layering of the material in the geosphere according to density. Since the hydrosphere is part of the geosphere, it is quite natural to regard its formation as a result of the same gradual and long-lasting differentiation of matter.

It is well known that magma contains a great deal of water, which is separated out during crystallization. Thus, for example, it has been reckoned that granitic magma contains up to 6 per cent of water by weight, whereas after crystallization only 1 per cent of water remains in the granite. It has been calculated that the granite layer underlying the continents, having an average thickness of about forty kilometers (the figure is greatly exaggerated), during the process of crystallization might have separated a sufficient amount of water to form the entire hydrosphere. Rubey has suggested a hypothesis to explain the filling of the oceans with water that envisages segregation of the granites in the process of differentiation, formation of granitic continents floating continually higher, and, parallel with this, separation of deep-seated water to fill the deepening ocean basins.

But in addition to the fact that the thickness of the granitic layer in Rubey's continents has been greatly exaggerated (about three times), it must be remembered that most of the granites forming the base of the continents are of Archean age, whereas the great increase in the volume of water in the oceans took place much later, during Mesozoic and Cenozoic times. It would be better to try to connect the separation of the water with the production of the basaltic lavas observed in the formation of the oceans.

This problem leads directly to the following no less difficult one: the most recent geophysical investigations have shown that the division of the earth's surface into continents and oceans is associated with the heterogeneous structure of the crust. The continents contain a granitic layer, and the Mohorovičić discontinuity lies at a depth of about 35 km. In the oceans there is a basaltic layer directly beneath the sedimentary

mantle, and the same discontinuity lies at a depth of 8 to 12 km below the water's surface. This structure of the crust is seen in all the oceans.

Although the notion of the secondary formation of the oceans on the site of former continents is far from new, and was widely held a few decades ago, many investigators have recently come to believe that this heterogeneity of the earth's crust eliminates the possibility that continents or shallow seas could have become oceans. In connection with this, the idea has gained currency that the ocean basins are primary structures and that the continents expanded gradually at the expense of the oceans as a result of the unequal differentiation of the material in the earth.

All that has just been said, this author believes, shows quite convincingly that the departure from the old view is not well founded and is, moreover, contradicted by the obvious geologic facts. These show that the deep oceans are relatively young, whereas the granitic layer of the earth's crust is predominantly older—of Archean age. Since continental masses were previously situated in the areas of present-day oceans, there is no justification for thinking that at that time there was no such granitic layer here as is now contained below the continents. Hence the explanation for the heterogeneity of the earth's crust must be sought not in the intensive granitization of certain parts of it but in a *basification*, or *basaltization*, of other parts, which took place after a time when the crust had been relatively homogeneous and "granitic."

Enormous basic volcanic extrusions following after the fracturing of the crust accompany the subsidence of the ocean basins. The present continents are fragments of much greater ancient continents; the angularity of their borders agrees much better with this interpretation than with the view that sees the continents as the result of an ever greater accumulation on the surface of sialic masses raised from the depths. One can imagine "solution" of the granitic layer by rising superheated basic magma; such an idea has come up from time to time among geologists and, in the opinion of many petrologists, is not inconceivable. This played a great role in the conception of subcrustal processes developed earlier by Daly (1933). The concrete details of this process are still not explained, but the paths to a solution of the problem may nevertheless be found in the process of basic metasomatism through the action upon the crust of superheated basic masses in the depths of the earth (Tikhomirov, 1958).

Everything that has been said in this chapter shows that, despite our poor knowledge of the geologic structure and history of the oceans, the division of the earth's surface into continents and oceans is a more fundamental process than its division into geosynclines and platforms. The vestiges of both geosynclines and platforms may be found, submerged and basalticized, at the bottom of the oceans.

CHAPTER 34

GENERAL SURVEY OF THE GEOTECTONIC PROCESS

The preceding chapters have already explained the relationship between the various kinds of tectonic movements, and it has been seen that all these movements are closely connected. Thus one may speak of a single geotectonic process or of a single process of development of the earth's structure, which manifests itself in different forms.

Three kinds of tectonic movement has been distinguished: oscillatory, folding, and faulting. The close connection between igneous activity and tectonic movements has also been mentioned.

Among the kinds of tectonic movement indicated, the most important is oscillatory movement. Oscillatory movements are ubiquitous and continuous, forming the background for the local and temporary appearance of the other tectonic processes, so that the latter also depend on the same fundamental vertical movements of the earth's crust.

It has been shown that the history of the structure of the earth's crust reveals, on the one hand, a general tendency in its development and, on the other hand, a periodic recurrence of tectonic processes superimposed on this general tendency. This recurrence is determined chiefly by the history of the oscillatory movements, which impart a definite rhythm to the whole process of tectogenesis and divide it into cycles or stages. The cycles are due to the periodic recurrence to be observed in the development of the oscillatory movements (general oscillations). Although the surface of the earth always contains both areas of uplift and areas of subsidence, upward or downward movement generally predominates at a given time. The length of each such primary cycle is about 150 million years on the average; the major cycles, however, are complicated by subordinate, secondary cycles.

It has been noted, further, that during the course of the major cycles the earth's crust is divided into wavelike oscillatory movements of greater or less intensity. The areas of greater intensity are called *geosynclines* and those of less intensity, *platforms*. Geosynclines have been characterized not only as areas of large-scale oscillatory movements but, most important, as areas of great contrast and amplitude in these movements. Platforms, on the other hand, are distinguished by the

small magnitude, contrast, and amplitude of these same movements. The magnitude and abruptness of the oscillatory movement in geosynclines, in turn, cause the great intensity and greater variety that distinguish all the other kinds of tectonic movement in these regions.

A schematic reconstruction of the overall course of a cycle in the history of the tectonic movements in a geosyncline is as follows:

The cycle in the development of a geosyncline has two stages; in both stages the geosyncline is divided into zones of subsidence and uplift. But subsidence predominates in the first stage and uplift in the second. The transition from the first stage to the second involves a general inversion of the geotectonic conditions.

The predominance of subsidence in the first stage leads to a gradual expansion of the zones of subsidence—the intrageosynclines—at the expense of the zones of uplift (the intrageanticlines) and the adjoining parts of the platforms. This results in marine transgression. There is great accumulation of sediments, with the sequences changing from terrigenous toward limestone. The earth's surface, in addition to the expansion of the marine basins, undergoes a simultaneous leveling of its topography; both these factors produce a change in climatic conditions in the direction of a more humid, warm, and uniform climate.

The other kinds of tectonic movement are combined with oscillatory movements according to observable laws. Folding is still not intensive: these are the preliminary phases of folding, confined to the intrageosynclines, and they produce only weak intermediate and discontinuous folds. Each such phase is accompanied by short-lived, sharp uplifts of the earth's crust into intrageanticlines. In certain cases, the folds formed in one place are replaced in others merely by uplifts. There is no folding at this time along the axis of the geosyncline; here the accumulation of sediments continues without angular unconformities, if not completely without interruption. Faulting is not typical of this stage. In particular cases, however, the wavelike oscillatory movements are accompanied by subsidiary deep-seated ruptures and become ruptural oscillatory movements. Various minor faults (thrusts, reverse faults, and normal faults) accompany the preliminary phases of folding. Igneous activity takes the form of extrusions (ophiolites) and sills, which are predominantly basic in composition, and, to less degree, of minor intrusions.

Inversion is the all-important moment of change in the life cycle of the geosyncline. There is not only a transition from predominant subsidence to predominant uplift but also a change in the locations of the areas of uplift and subsidence. A region of uplift arises within the intrageosyncline, whereas subsidence (in the marginal deeps) continues on the periphery; as it gradually rises, the central uplift turns the intrageosyncline into a zone of upward swelling of the earth's crust, and the marginal basin, in moving outward, becomes involved in the

neighboring geanticlines and turns them completely or partly into intermontane basins. The edges of the platforms are also drawn into the subsidence, forming foredeeps. Zones of uplift are replaced by zones of subsidence, and vice versa, according to the partial or local inversions within the geosyncline. The process begins within the intrageosyncline, and extensive intrageanticlines may be partially preserved as zones of uplift to the end of the cycle.

Later the central uplift expands, the intermontane basins decrease in extent, the foredeeps migrate outward, upwarping becomes more and more predominant, and the cycle ends with a general archlike uplift of the area (the stage of mountain building). Troughs and large interior basins are superimposed on this uplift.

In accordance with the change in the relationship between uplift and subsidence of the topography, there is a change in the sedimentary sequences from upper terrigenous to lagoonal, molasse, and interior basin formations. The climate becomes more continental, the extremes of climate become greater, and glaciated areas appear.

The chief phases of folding are associated with the period of inversion. They are confined to a newly rising central uplift and fill the geosyncline from the axis of the intrageosyncline, moving outward as the central uplift expands. The concluding phases bring the intermontane and frontal flexures into the folding. When internal basins are formed, the late phases of folding do not appear in them.

It can also be seen that the type of folding, the intensity, the lateral extent, and the slope depend closely on the oscillatory movements. The trend of the folds is fundamentally parallel to the isopach lines; the movement of the rock masses is directed outward from the intrageosyncline toward the intrageanticlines or—what amounts to the same thing—from the central uplift toward the intermontane basin and foredeeps, and in the marginal uplifts it is from the interior basin outward, again toward the intermontane or marginal basin. The intensity of the folding is connected with the amplitude of the oscillatory movements and with the rate of change in the thickness of the sedimentary cover. The most intensive folds are formed in the area of the intrageosyncline; toward the intermontane basin and the foredeep the folds become gentler, and they are replaced by intermediate and discontinuous folding. The folding is accompanied by thrusting and, to less extent, by faulting. Each phase of folding is connected with a temporary increase in the rate of uplift of the earth's crust.

During the second half of the cycle, the upward swelling of the crust in the central uplift, which involves tension of the crust, causes fractures to appear. At first these are bending fractures, which then grow into faults. This fracturing of the crust characterizes the period of mountain building.

Before the inversion, the early igneous activity dies out; but from

the moment the inversion begins, igneous activity breaks out with new force and in new forms: the main phases are accompanied by the formation of granitic batholiths. Later, as the crust begins to be fractured, fissure intrusives and veins are formed and the formation of faulted fractures ushers in the extrusive activity of the period of mountain building.

If the above explanation of the origin of folding is accepted as valid, this conception of mountain building reveals the intrinsic unity and the single cause of the entire processes. This cause is the following: immediately before the inversion, there is an increase in the forces of uplift in the subcrustal regions of the earth. The action of these forces on the crust produces the whole complex of tectonic phenomena in succession, including the granitic intrusions, the uplift and the flow of matter leading to folding, mountain building, fracturing of the crust, and extrusion of lava upon the surface. When large quantities of magma are expelled, there is a tremendous loss of energy, which forms the background of the increasing subsidence that begins the next cycle. Since all the forms of tectonic movement in this succession, from the period of inversion to that of mountain building, are the result of the same forces of uplift, these forms of movement are, as it were, interchangeable. Thus uplift and folding may be replaced by fracturing and the pouring out of magma on the surface. Figure 116 shows a diagram of the development of a geosyncline and of its final structure.

It has been shown that the geosynclines of succeeding cycles are formed on the site of the geosynclines of previous cycles; the later intrageosynclines arise on the basins, intermontane and foredeep, that existed at the end of the preceding cycle. But during at least the previous three cycles (and more probably four or five), only part of the geosyncline of the preceding cycle became a geosyncline in the next cycle. Part of the area ceased to be a geosyncline, so that from cycle to cycle there has been a step-by-step growth of the platforms and diminution of the geosynclines.

A projection of this process backward in the history of the earth suggests that geosynclinal conditions originally encompassed, if not the entire planet, at least the greater part of its surface. Thereafter there was a growth of stable conditions, spreading outward from certain centers whose position is regular and possesses a very rough symmetry. The increase in the area of the platforms and their rounded shape is the reason that the geosynclines of later cycles take the form of narrow belts winding sinuously between extensive platforms; in addition, these belts basically follow the directions of lines of latitude and longitude. As their area decreases, the geosynclines tend to form separate ovals. The platforms that have been formed over earlier geosynclines follow their distribution of movements and, in general, repeat the complex of movements characteristic of the geosynclines, but with a lower intensity.

The subgeosynclines and subgeanticlines into which the platform is divided at first more or less follow the positions of the areas of subsidence and uplift at the end of the cycle in the geosyncline that last occupied this region. Later on, however, this repetition is disrupted: the areas of subsidence and uplift on ancient platforms are typically much more extensive and rounded than in geosynclines. These far-reaching subsidences and uplifts extend into the areas of the adjoining geosynclines and here produce general movements, in contrast with the more localized and abrupt movements that properly characterize geosynclines. The geosynclinal movements are, in a manner of speaking, a surficial phenomenon superimposed on platform movements, which thus give the impression of having a more deep-seated origin.

Often extensive areas of subsidence appear on platforms immediately after the latter have been formed; in zones that have ended their geosynclinal development and made the transition to platform conditions there are sometimes interior basins—broad, gentle subgeosynclinal depressions lying unconformably upon the former geosyncline. An example of such a depression is the western Siberian lowland.

Between geosynclines and typical platforms there is an intermediate state, that of the parageosyncline. Parageosynclines are semiplatforms; within them the local inversion is weaker or fails to take place. Folding of the intermediate type is typical. Parageosynclines are formed within geosynclines, marking a gradual transition to platform conditions, and they also occupy large areas outside geosynclines.

There are also two stages in the history of platforms during a cycle of development. The platform generally undergoes subsidence during the first stage and uplift during the second. Subgeosynclines and subgeanticlines suffer certain displacements and changes in their outlines, but radical redistribution and complete inversion do not take place. The great overall oscillations of the platforms are synchronized with, or lag somewhat behind, the oscillations in the geosynclines during the same cycle. Subsidence of the platform is, in a number of cases, accompanied by extrusions of magma in the subgeosynclines.

Although no local inversions in the full sense take place on the platform, there is an analogous but much less marked change involving the formation of discontinuous folds in the subgeosynclines. An analogy has been drawn in principle between these and the central uplifts supported by the fact that the formation of the discontinuous folds is, in places, accompanied by intrusions; these are minor intrusions, however, rather than batholiths.

As with geosynclines, formation of uplifts on platforms, represented by discontinuous folds and anteklises, is accompanied by fracturing of the earth's crust and formation of bending fissures and normal faults. Fissure intrusions and extrusions are encountered along with the faults on the uplifts.

An important feature of the tectonic process in general is the tendency of uplift to replace subsidence and of subsidence to replace uplift. The former is reflected in the regularly recurring formation of central uplifts within intrageosynclines and of discontinuous folds in subgeosynclines, and the latter is reflected in the formation of superimposed troughs with their typical lignite beds and zones of subsidence on subgeanticlines and discontinuous uplifts.

It has been noted that the periodic recurrence in the tectonic process is a secondary phenomenon superimposed on the general tendency of its development. This overall tendency, since the end of the Precambrian, has been an expansion of the platforms at the expense of the geosynclines. This has been more rapid in the southern hemisphere, where the platforms became very large and the geosynclines were contracted earlier than in the northern hemisphere.

Other irreversible changes may be seen by comparing the processes that transformed the structure of the earth's crust in the early Precambrian to post-Proterozoic tectonic processes. The linearity of the zones of tectonic development that characterizes the geosynclines of later times is absent in the early Precambrian. In Archean times the deformations took the form of very large domes, whose origin is closely connected with the intrusion of granitic batholiths. On the flanks of these domes the strata are often vertical or even overturned; they are complicated by small disharmonic folds, which often have vertical axes. These features, along with widespread scattered injections and migmatites, and the general plasticity of the rocks, as reflected in small complex disharmonic folds, indicate the great plasticity of the earth's crust and the material of which it is composed.

The tendency of the zones of tectonic activity to assume a rounded oval shape is partially maintained in post-Archean times as well. At various times and in various places, however, this has yielded to the formation of elongated linear zones of uplift and subsidence, which have frequently maintained a uniform trend over a great distance. This linearity has been attributed to the appearance of a network of basic deep fractures in the earth's crust. Moreover, their effect on the positions and trends of the zones of uplift and subsidence is manifested in the relatively intensive movements that have taken place in geosynclines, parageosynclines, and young platforms (in Western Europe outside the area of the Alps, in the central Kazakhstan of Hercynian times, etc.). On ancient platforms, where the vertical movement is considerably diminished, one again observes a predominance of rounded or oval randomly located areas of uplift and subsidence; the subgeanticlines and subgeosynclines of the Russian platform are examples.

The general tendency in the history of the earth's crust toward a clamping of its movements was disrupted in the middle of the Tertiary,

when there was a renewal of activity on the platforms and a new formation of sharp intensive uplifts and areas of subsidence in certain places, primarily in central Asia. Such reactivated regions have much in common with geosynclines, but they can scarcely be considered as the equivalent of actual preplatform geosynclines. Probably this is a new form of development of the earth's crust, which follows the formation of platforms.

During the course of the tectonic process, the structure of the earth's crust becomes more complex from one cycle to another as a result of the deformations produced by folding. The structure of the crust is also affected by changes in the structures of the rocks and by the formation of various tectonites. Igneous activity, particularly intrusive, plays the most important role in the irreversible development of the earth's crust. Granitic intrusions fill up the crust with ever-increasing amounts of granitic material and thus cause the gradual growth of the granitic layer. All these changes generally lead, apparently, to a diminution of the plasticity of the earth's crust and an increase in its brittleness. Thus the reaction of the crust to successive tectonic forces shows an increasing dominance of rupture over plastic deformation. The combination of a general trend with a periodic recurrence gives the entire process of tectogenesis the appearance of a spiral.

The division of the earth's surface into continents and oceans belongs to a special order of phenomena. It has been suggested that the oceans were formed fairly recently. In this view there were no oceans at the beginning of the Paleozoic; they were formed on the sites of continents and shallow seas and then expanded at the expense of the latter. The ocean boundaries do not match the boundaries between platforms and geosynclines; both can apparently be traced on the continents and in the oceans.

It is interesting to note that the oceans were formed chiefly in the Southern Hemisphere, where great uplifted platforms were also formed earlier than in the Northern.

The formation and growth of the ocean basins, the increase in the total area of the platforms, and their transition to newly reactivated forms of development similar to those seen in central Asia are all part of the general tendency in the history of the earth's crust.

It must be said, however, that the processes that control this overall tendency in tectogenesis have as yet been relatively little studied, and the laws governing them are not sufficiently well known. The most important task of geotectonics is the discovery of these laws and the study of the fundamental continuity in the development of the earth's crust throughout all the geotectonic cycles and during the entire course of geologic time.

CHAPTER 35

PRINCIPLES OF GEOTECTONIC REGIONALIZATION

GENERAL REMARKS

It is clear from what has been said previously that different parts of the earth's crust have their own peculiar histories of tectonic development. These differences in tectonic history result in corresponding differences in the composition of the rocks (formations), thickness, structure, and degree of metamorphism. The differences in development, moreover, lead to the formation of different complexes of ore minerals.

Thus there arises the task of dividing the surface of the earth into various structural zones—of geotectonic regionalization. This will result in a map or diagram showing each of the different portions characterized by its own peculiar structure and history.

Since tectonic movements affect the composition and conditions of deposition of sedimentary strata (sedimentary formations) and are accompanied by various kinds of igneous activity, all aspects of geologic structure can and should be reflected on diagrams of geotectonic regions. The ages of the structures in question must also be noted in geotectonic regionalization. Thus geotectonic regionalization is essentially the classification of various parts of the earth's crust according to their structures and structural histories.

Like any general classification of natural phenomena, such a categorization will be to some extent tentative. There are no sharp boundaries between different conditions in nature; one set of conditions replaces another through a series of transitions in the border areas. We have already encountered the transition from the structural conditions of a folded zone to those of a platform and seen the necessity of indicating a certain intermediate zone, containing features of both the folded zone and the platform in various relationships to each other.

Every classification nevertheless presupposes the existence of distinct boundaries between the phenomena belonging to the different classes.

Such boundaries are almost always artificial and contradict the natural continuity of the transitions. Nonetheless, all branches of knowledge have developed systems of classification as a necessary step in scientific investigation; these are widely used both in theory and in practice.

The classification of geotectonic zones is of great importance in practical geology as well as in theory: the formation of mineral deposits, as explained above, is closely associated with structural developments. A correctly drawn geotectonic diagram may therefore be the basis for predicting the distribution of various groups of ore minerals.

In every scientific classification there is a principle that forms the basis for dividing observed phenomena into classes and for the degree of detail in the subdivision. The most important requirement in scientific classification is the consistent maintenance of the principle by which the phenomena are classified. This requirement is often not met, out of considerations of practical convenience, but such deviations must not be carried too far.

What are the principles by which the crust of the earth can be subdivided into geotectonic regions? Morphologic principles are the simplest; in this case the subdivision will be made exclusively on the basis of the structural form. The two broadest categories might distinguish those portions of the surface (disregarding the mantle of Quaternary sediments, of course) in which the rocks have been crumpled into continuous folds from those in which they are flat-lying or form only discontinuous folds.

By this principle, for example, one might distinguish such structural regions as the Fennoscandian shield, which contains outcrops of rocks that have been tightly crumpled into folds, and the Russian plain, in which the rocks at the surface are flat or very gently inclined. The boundary between these regions will be the southeastern border of the Baltic shield, where the outcrops of Precambrian folded rocks are replaced by outcrops of flat-lying Paleozoic sediments.

A more detailed classification can be made in distinguishing dislocated portions of the crust, by subdividing such portions in turn into several categories. Stille (1918), for example, distinguished the following:

1. Blanket structures, characterized by the predominance of huge tectonic thrust sheets over other types of dislocations.
2. Folded structures, in which folds are the chief structural forms.
3. Faulted fold structures, in which the folds are broken by faults.
4. Block-faulted structures, or areas of which the structure is determined by the vertical displacement of individual blocks separated by faults.

V. A. Obruchev (1935–1938) added to this list:

5. Folded-block structures, in which the faults break up areas that have previously been subjected to intensive folding.

To achieve greater detail, one may distinguish individual folds, thrust sheets, faults, and even separate parts of these dislocations, which may then be shown on the map. The details of the method of constructing such structural maps are described in courses in structural geology. One can easily see, however, that morphology is not an adequate basis for distinguishing structural zones and cannot produce a clear picture, because of the fact that the dislocations will be of different ages.

Let us return to the last-cited example. It is clear to anyone with a knowledge of the fundamentals of general and regional geology that the boundary between the crystalline rocks of the Fennoscandian area, which have been squeezed into folds, and the flat-lying Paleozoic sediments of the area around the Baltic Sea and Lake Onega and of the Onega-Dvina interstream divide is not important enough to be one of the main boundaries on a geotectonic map. The crumpled crystalline rocks do not come to an end at the point where the Paleozoic sediments begin on the surface but extend beneath them throughout the whole area of the Russian plain. We are not dealing with two adjacent portions of the earth's crust but with two levels of the crust—two complexes of rock lying one above the other, with an erosional interval between them. It is most probable that at one time the flat-lying Paleozoic rocks extended over the folded Precambrian considerably farther to the north-west than at present.

This example shows that in making a morphologic tectonic classification one must take the age of the dislocation into account, so as not to group folded zones of different ages in one category. Thus the principle underlying the classification becomes partly morphologic and partly historical. This is fully permissible, however, from the standpoint of a correct classification, since in this case two bases (age and morphology) are required at different levels of subdivision. Two operations have been carried out in succession: first, areas of folded and flat-lying beds were distinguished in the earth's crust; then, areas with folding of different ages were segregated within the common area of folded structures.

If a very detailed classification is required, these regions may be subdivided into still smaller structural elements. Within a folded zone these might first include anticlinoria and synclinoria, then folds of various orders and types, and finally other subsidiary structural elements. On a platform one might distinguish anteklises and syneklises, faulted folds, etc. Thus, after subdividing on the basis of morphology and age, we return again to a morphologic basis for categorization. Such a geotectonic regionalization results in a diagram showing the distribution of folded zones and platforms of different ages—that is, of zones associated with different cycles.

It follows from the above that, even without much detail, a complete

geotectonic regionalization for the Hercynian and older cycles demands meticulous effort and, especially, a large amount of data. The amount of such data presently available is far from sufficient; thus geotectonic maps showing the extent and the individual elements of the internal structure of the Hercynian folded zones, with a full division of the Hercynian platforms into synclises and anteklises and a delineation of the faulted uplifts, can now be drawn only for a few areas. Geotectonic diagrams of large territories can be drawn only approximately and to a small scale.

One method of approximation in delimiting geotectonic regions can be used at the present time. First, the zones of Alpine folding are picked out; then, within the Alpine platforms, the portions lying on folded bases of different ages—Hercynian, Caledonian, and Precambrian—are indicated. Thus the zones of folding older than the Alpine are shown on the map not throughout their entire extent, but only where they form outcrops beyond the younger folded zones and lie beneath younger platforms. Tectonic maps of the U.S.S.R. and of the whole world were constructed on this principle by A. D. Arkhangel'skiy (1941) and later writers.

Since later geosynclines are situated over earlier ones (the exceptions to this rule, which can be seen at the edges of certain geosynclines in the migration of the foredeeps, being of little importance), it may be assumed with sufficient accuracy that folded zones of earlier cycles existed on the site of the folded zone of a later cycle. Thus wherever younger folded dislocations cover other folded zones on the map, the older folded zones may be considered to continue beneath them. But the internal subdivisions of the folded zones may be shown only in those areas where the given folded zone is the youngest. Only the zone of Alpine folding can be shown throughout its entire extent. Anteklises and synclises of Alpine age may also be indicated on the same map.

The methods described above will give some idea of the distribution of various tectonic conditions over the earth's surface; this, however, will reveal only the morphology, or geometry, of the structures. The picture still contains no indications of the relative nature of the rocks, of the sedimentary formations in various places, of igneous activity, and of metallic and nonmetallic mineral deposits. Since this deficiency of the method of geotectonic regionalization seriously affects the results of practical work in geology, the means must be found to overcome it. Only by basing the system on data that reveal the material composition of the rocks can one meet the demands that are legitimately made on the methods of geotectonic regionalization.

Numerous attempts have been made to satisfy this requirement by subdividing individual regions according to kinds of geologic evidence that have purely local significance, without trying to maintain the

same basis for subdivision throughout. In these cases a given territory is divided into areas distinguished by differences in composition of the rocks, ages of the outcropping rocks, thicknesses, structures, kinds of igneous activity, and other characteristic features. One area will be delimited to a greater degree according to one feature and another area according to another feature.

One example is the geologic subdivision of the Caucasus proposed by V. P. Rengarten (1930). Some of the subdivisions in his scheme are geographic, and the names of others reflect the structure: for instance, the geographic zone north of the Caucasus is shown as a unit equivalent in importance to the Transcaucasian zone of gentle folding.

Similar schemes of geologic subdivision have been suggested for many other areas, both of the U.S.S.R. and of other parts of the world, such as the subdivision of the Siberian platform by N. S. Shatskiy (1932), which was also made according to various local criteria.

It must be admitted that geologic subdivisions of this nature are often very helpful in practical work; they emphasize the most important and useful features of a particular area and produce an overall picture from geologic data of various kinds. Such schemes, however, give the distribution of various geologic features without showing the relationships between them, their origin, or the laws governing their disposition. They merely bring together all the known data and serve as a sort of aid to memory. But they do not suggest any explanations of the data or any interpretations or predictions of the geologic conditions that exist beyond the limits of the known area.

SUBDIVISION INTO REGIONS ACCORDING TO DIFFERENT TECTONIC DEVELOPMENT

We shall attempt to set forth some principles of geotectonic regionalization that may form the basis of a system meeting, if not all, at least most of the requirements made of it. Each separate zone, on the one hand, should be characterized by rocks of specified composition and structures. On the other hand, the subdivision into structural zones must have some consistent, universal basis, making it possible to draw general conclusions regarding the history of a given area, the relationships between different zones, the rules governing their disposition, and the extension of a given set of geologic conditions beyond the limits of the area that has been studied directly.

In principle, it is possible to solve this problem, since it is known that the various kinds of tectonic movements are closely interrelated. A specific succession of oscillatory movements is associated with definite types of folding and igneous activity, and since the oscillatory movements determine the facies and thicknesses of the sedimentary deposits,

a specific facies, or, more properly, the development of specific sequences, should correspond to a specific structure or igneous formation.

On the basis of this interrelationship between various tectonic processes, one may distinguish not different types of structures but different types of tectonic development, including all the various kinds of tectonic movement, and thus subdivide the earth's surface according to the distribution of these types of development.

First of all, as before, one must distinguish geosynclines and platforms of various ages. Parageosynclines may also be noted as areas of intermediate development.

Geosynclines, on the basis of oscillatory movements, were defined above as zones in which these movements are more abrupt and show greater contrast. Taking a more complex view of geosynclines, we may add to the first feature the fact that the development of a geosyncline is characterized by general and local inversions, by a definite succession and disposition of sedimentary formations, by the development of predominantly continuous folding, and by intensive igneous activity resulting especially in the intrusion of batholiths. It must be stated categorically that this combination of features consists not of disconnected phenomena independent of each other, which may also occur separately, but of closely interdependent characteristics.

Platforms are characterized by little contrast and very gentle oscillatory movement, by the absence of complete inversion in their development, by a less complete succession of sedimentary formations than in geosynclines, by discontinuous folding, and, as a rule, by less igneous activity. All these features of the development are also interrelated.

In parageosynclines the amplitude of the oscillatory movements is less than that usually observed in geosynclines and greater than the amplitudes typical of platforms. As on platforms, local inversions either are lacking or are reflected in the uplift of isolated individual folds following isolated local subsidence. In other cases, there is local inversion, but never with any considerable amount of uplift. The folding, as a rule, is intermediate—ridge-shaped, box-shaped, and often discontinuous. The folds may or may not be inclined. Intrusive igneous activity takes the form of minor intrusions and stocks that are most often of alkaline composition.

A more detailed subdivision of geosynclines may be made according to the zones into which the geosyncline is divided at the end of its cycle of development: these include foredeeps and intermontane basins, central uplifts, stable geanticlines, and epigeosynclines. The development of each such zone will have its own specific features, reflected in the composition of the rocks and in the structure. The characteristic features of each of the enumerated zones will be described briefly below (see Fig. 116).

Central uplifts are formed from intrageosynclines by inversion. The crest of the uplift is composed of the preinversion lower terrigenous and limestone sequences. These strata show signs of being intensely crumpled during the first main phase; the folds on both sides of the axial plane of the uplift are often asymmetrical. There are many overthrusts, and very large normal faults are usually produced at the end of the cycle. There are sills of basic composition and primarily basic blanket flows that occur conformably with the sedimentary strata. There are also lava flows, which are the youngest in the cycle and whose composition is mainly of the Atlantic type; these are clearly associated with and produced during the formation of the mountain topography at the end of the cycle. Their occurrence is unconformable with that of the entire complex of crumpled rocks.

The dislocated rock complex contains younger intrusions, granite batholiths, and the complicated series of fissure intrusives associated with them. The rocks undergo regional metamorphism, which is so intensive in places that they are altered beyond recognition and appear considerably older than their actual age. If there has been any great amount of uplift after the inversion, most of the preinversion rock complex of the given cycle may be eroded away, allowing the rocks of older cycles to crop out at the surface. An example is the ancient granites of the main Caucasus range.

In its overall structure the central uplift is an inverted anticlinorium. The preinversion rocks of the central uplift are distinguished by their great variety of ore deposits. Some of these are associated with the basic sills and form accumulations of heavy-metal ores: platinum, nickel, chromium, and iron in the form of magnetite. These deposits are found within the igneous bodies. Pyritic deposits of copper and other metals are associated with the minor intrusives of the preinversion stage.

Another group of ore deposits is associated with the granitic intrusions and the vein and fissure intrusions that accompany them. A large variety of metals may be found in these: gold, copper, lead, silver, tungsten, molybdenum, tin, bismuth, antimony, mercury, and many others. Deposits of these ores are found in the exterior portions of the batholiths and in the contact zones of fissure intrusions and veins.

The central uplift, as it expands, often encompasses adjacent zones where postinversion sequences have accumulated—the marginal basins. The structures of the latter are thus partially or fully incorporated into that of the central uplift.

The flanks of the central uplift, moreover, characteristically contain thick accumulations of upper terrigenous sequences. In the zone where the latter are extensively developed, the preinversion sequences have small thicknesses.

The upper terrigenous sequence is usually warped into large broad

similar folds inclined outward, toward the adjacent basins; the continuous folding becomes intermediate as one proceeds in this direction. Overthrusts and overturned folds are extensively developed; this folding corresponds in age to the later main phase of folding.

Igneous bodies are not typical of these formations on the flanks of the central uplift, which are penetrated by fissure intrusions and veins from the central zone. Minor intrusives (stocks, laccoliths) are frequently encountered, and the degree of metamorphism of the rocks is slight. Mineral deposits are associated either with the sills in the pre-inversion strata, if there are any such, or with the fissure intrusions and veins. Subsequences composed of caustobiolithic rocks contain coal beds and bituminous shales; oil is less frequently encountered.

Examples: The Greater Caucasus, the Kara-Tau, the Hyssar Range, Wales, the central zone of the eastern Alps, and many others.

Foredeeps and intermontane basins are characterized by thick accumulations of lagoonal and molasse formations produced late in the given cycle. The caustobiolithic rocks, and sometimes part of the flysch sequence, extend as far as the deeps. Both preliminary and terminal phases appear in the intermontane basins and terminal phases in the foredeeps; the unconformities that represent the main phases are often not observed. The folding is less severe and of intermediate and discontinuous type, and there are extensive diapirs. Igneous activity is not great: certain basins typically contain minor alkaline intrusives. Extrusive volcanic activity is sometimes seen. There is practically no metamorphism. Faults, both normal and reverse, are associated with individual discontinuous and intermediate uplifts.

Typical mineral deposits include gypsum, anhydrite, halite and sylvite, oil, and coal (in smaller amounts). Some metallic and nonmetallic minerals are associated with the minor intrusives; the metallic deposits include some rare elements (zirconium, tantalum, niobium, etc.), and semimetals. Cupriferous sandstones are common in the molasse formations.

Examples: The Ural foothills, the Kura-Rion depression, the Tadzhik basin, the molasse zone of the Alps, and many others.

Stable intrageanticlines are the remnants of preinversion intrageanticlines that have not become intermontane basins. Here the sediments are thinner, and particular strata may be missing from the section. There are manifestations of the less intense preliminary phases of folding, accompanied by igneous activity (minor intrusions, fissure intrusives and extrusives). Numerous abrupt normal faults appear. Mineral deposits are associated with the intrusives. The overall structure of the stable intrageanticlines is that of noninverted anticlinoria. Extensive stable intrageanticlines approach platforms in structure (as in the case of the Hungarian intrageanticline).

Examples: The Miskhan zone of the Transcaucasus, the Ural-Tau, the Great Basin of the Rocky Mountains, and others.

Epigeosynclines are formed on central uplifts and on stable intra-geanticlines; the lignite-bearing basins on the eastern slopes of the Urals are examples.

Two series make up the section of an epigeosyncline: an upper series containing the epigeosynclinal sequences and a lower, more complex series. The upper series is undisturbed and of the same age as the very youngest sediments of the cycle. The folds are poorly developed and discontinuous. This complex contains only sills and a very few alkaline minor intrusives; there may be extensive volcanic activity on the surface, however, manifested in the formation of lavas with compositions of the Atlantic type. The mineral deposits in this series include coal (mainly lacustrine), natural gas, and oil.

There are an angular unconformity and an erosional interval between the epigeosynclinal and the lower series; this unconformity and erosional interval are naturally greater over the central uplift than in places where the epigeosyncline covers an intermontane flexure. The composition and structure of the lower series are the same as those described above in the case of the central uplift and the stable intra-geanticline.

Parageosynclines occupy an intermediate position between typical geosynclines and typical platforms. Local inversions are frequently observed in them. They may have cross sections of various shapes, depending on their position in relation to the geosynclines and platforms. In the Rion parageosyncline, for example, preinversion sediments are overlain by intermontane basin deposits. In other cases, the succession and disposition of the sequences more closely resemble platform conditions: especially in the normal cyclic succession of sediments from terrigenous to carbonate and again to terrigenous, as the area of sedimentary accumulation increases in extent and then decreases again. As the adjacent central uplift expands and develops, the parageosynclines are often incorporated into it; this has taken place in the northern Caucasus.

Parageosynclines situated within or next to geosynclines contain extensive limestone deposits, as has been observed in the Rion parageosyncline in the Caucasus. This testifies to the small amplitude of the oscillatory movements and the absence of any great uplifts that might provide a source of clastic material. In other cases, in which the parageosyncline is far removed from a geosyncline, it will contain accumulations of primarily continental deposits (as in China, for example), but these will show little variety, again testifying to the small amplitude of the oscillatory movements. The folding in parageosynclines is intermediate and discontinuous; their structure is that of a synclinorium.

Igneous activity is most often represented by minor alkaline intrusives, although extrusive igneous activity may be very intense. There

is a great variety of mineral deposits: coal beds, especially lacustrine, are extensive, and a complex of ore deposits is associated with the minor intrusives.

Platforms are first of all divided into portions that overlie folded basements of various ages. They are then further divided into subgeosynclines and subgeanticlines, the former having the structure of a syncline and the latter that of an antecline.

Subgeosynclines contain the most complete section of deposits, which are thickest in the place of greatest subsidence and always become thinner toward the outer edges of the subgeosyncline. Thus the deeper strata of a syncline show a greater amount of structural subsidence than the upper strata. The positions of the zones of greatest thickness may not be the same, however, for beds of different ages. Subgeosynclines are complicated by discontinuous folds of various types; the largest of these may have associated minor intrusives and fractures, such as fissures produced by bending, normal, and reverse faults. Moreover, subgeosynclines sometimes contain extrusives formed at the time of their subsidence. In subgeanticlines the thicknesses decrease toward the center, and many beds may be missing altogether. There are widespread fissures and normal faults, which in places have caused the formation of large grabens. Extrusives of Atlantic composition are frequently associated with the faults.

Platform deposits form the upper series of rocks, separated by a sharp angular unconformity from the lower, geosynclinal series that has been folded, intruded, metamorphosed, and subjected to other manifestations of geosynclinal tectogenesis. The upper series as a rule contains nonmetallic deposits: coal and oil in the appropriate sequences (lower and upper terrigenous), etc. The exception is bauxite, which sometimes is found in large deposits on platforms. Rare and minor elements may be associated with the minor alkaline intrusives (such as apatite, zirconium, tantalum, and niobium, etc.), whereas diamond, magnetite, copper, and nickel will occur in the basic intrusives of the traprock sequences. Another characteristic feature of subgeosynclines is the presence of interior basins.

The above system of subdividing geosynclines and platforms may be carried out in greater detail by distinguishing even smaller subzones, depending on the scale of the map. The zones specified in the system are characterized by definite structural features, on the one hand, and a definite historical development, on the other. Their positions in regard to each other follow a regular plan, and the relationships between them make it possible in many cases to foresee their disposition in areas that have been only partly studied.

Most likely, in applying this system to individual cases, obstacles will be encountered that may be connected with local complications or

deviations from the normal conditions. Cycles of higher orders, for example, may be superimposed on the basic cycle so that the succession of sequences will turn out to be more complicated. When there are local inversions at different times in different intrageosynclines of a complex geosyncline, these may exert a mutual influence on each other in regard to the succession of sequences. The geosyncline may contain intrageosynclines with different intensities of movement, as well as parageosynclines in which the conditions are intermediate. Certain metallic mineral deposits may be associated not with the latest but with earlier tectonic and magmatic cycles.

Everything said previously in regard to the method of revealing the structures produced by more ancient cycles in a step-by-step removal of the younger ones and in regard to the possibility of directly contouring older geosynclines only in the areas of later platforms (with the exception of the Alpine geosyncline) remains in force in the application of this system of geotectonic regionalization.

In conclusion, we may turn once more to Fig. 116, which illustrates some of the developments that may be observed in geosynclines.

CHAPTER 36

THE PRESENT TECTONIC STRUCTURE OF THE EARTH

Since a detailed discussion of regional tectonics is beyond the scope of this work, this chapter will present only a brief outline of the present structure of the earth, pointing out the basic elements in it. A regional subdivision will be made only in the case of the Alpine cycle. The Alpine geosynclines and platforms will be differentiated according to the principles established earlier in this book. Within the platforms a further distinction will be made between folded basements of various ages; their structures will be described briefly. (See the "Tectonic Diagram of the Earth" at the end of this book.)

EUROPE

The outlines of the Alpine geosyncline in Europe, as well as its differentiation into ovals, are already familiar. The most familiar part of the Alpine geosyncline in Western Europe is the western Alps, or the Maritime Alps, which are linked with the Apennines (Richter, 1939).

The Maritime Alps and northern Apennines (Fig. 305) are parts of a geosyncline developed during the Mesozoic and Cenozoic. It originated in the beginning of the Triassic, and its development ended in the Quaternary. On one side it touches the central massif of France, and on the other the platform of the Adriatic Sea and the plain of Lombardy, which is itself a narrow northerly offshoot of the great African platform, wedged between the Mediterranean and Hungarian ovals of the Alpine geosyncline (see Fig. 113).

A study of the facies and thicknesses of the Jurassic and Triassic deposits leads to the following subdivision of this geosyncline (Fig. 306): its western margin is occupied by the subalpine parageosyncline, which is in turn split into two intrageosynclines of a lower order by an uplift of *central granite massifs*, which are not everywhere well developed. This last subdivision is omitted here. In the subalpine parageosyncline

the Mesozoic is represented chiefly by limestones of considerable thickness.

In the east the subalpine zone borders on the Briançon intrageanticline, characterized by thin Mesozoic deposits, shallow-water facies, and the absence of a number of formations from the section. Still farther east lies the central intrageosyncline, called the *Pennine* on the French and the *Ligurian* on the Italian side. These two names designate the western and eastern margins of the same intrageosyncline, respectively; in this book the name "Pennine" will be used. Its Mesozoic deposits are represented chiefly by the *schiste lustré* facies, although the Upper

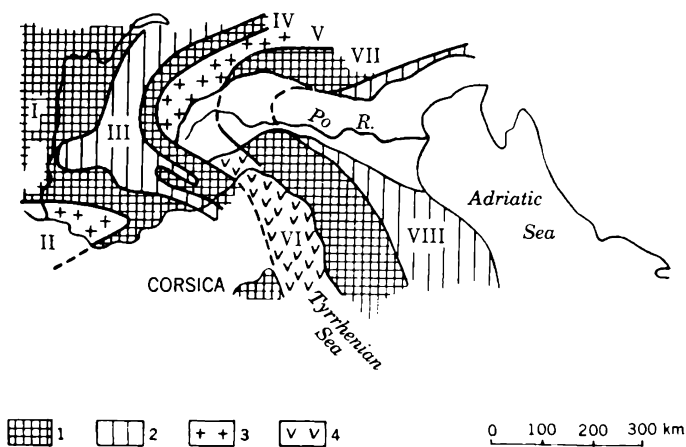


FIG. 305. Diagram of the structure of the western Alps and Apennine geosyncline: (1) platforms and intrageanticlines; (2) paraeosynclines; (3) intrageosynclines developed into central uplifts; (4) parts of intrageosynclines developed into interior depressions; (I) platform of the central plateau; (II) Pyrenees geosyncline; (III) subalpine paraeosyncline; (IV) Briançon intrageanticline and marginal uplift; (V) Pennine intrageosyncline and central uplift; (VI) interior depression on the site of the Pennine intrageosyncline; (VII) Tuscan intrageanticline and marginal uplift; (VIII) Lombard paraeosyncline.

Cretaceous is made up of limestones. The axis of this intrageosyncline passes through Genoa. East of the Pennine intrageosyncline is another intrageanticline—the Tuscan IGA. Finally, at the edge of the geosyncline, lies the paraeosyncline of Lombardy.

An examination of the Triassic facies and thicknesses leads to the conclusion that there was no Briançon intrageanticline during this period: instead, there was a "hinge" between the gently downwarped subalpine paraeosyncline and the more intensely downwarped Pennine intrageosyncline. The Briançon intrageanticline began its uplift at the beginning of the Jurassic. This is one instance of the structure of an early geosynclinal stage of development being complicated by the appearance of new zones within it.

Throughout the Mesozoic, the distribution of the zones within the geosyncline was preserved, with the geosyncline evolving toward a greater subsidence of the intrageosynclines, their expansion at the expense of the intrageanticlines, and a marine transgression. During the transition from the Early to the Late Cretaceous, this evolution was

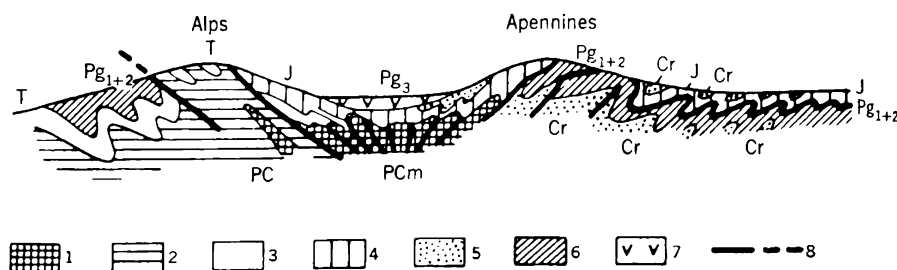
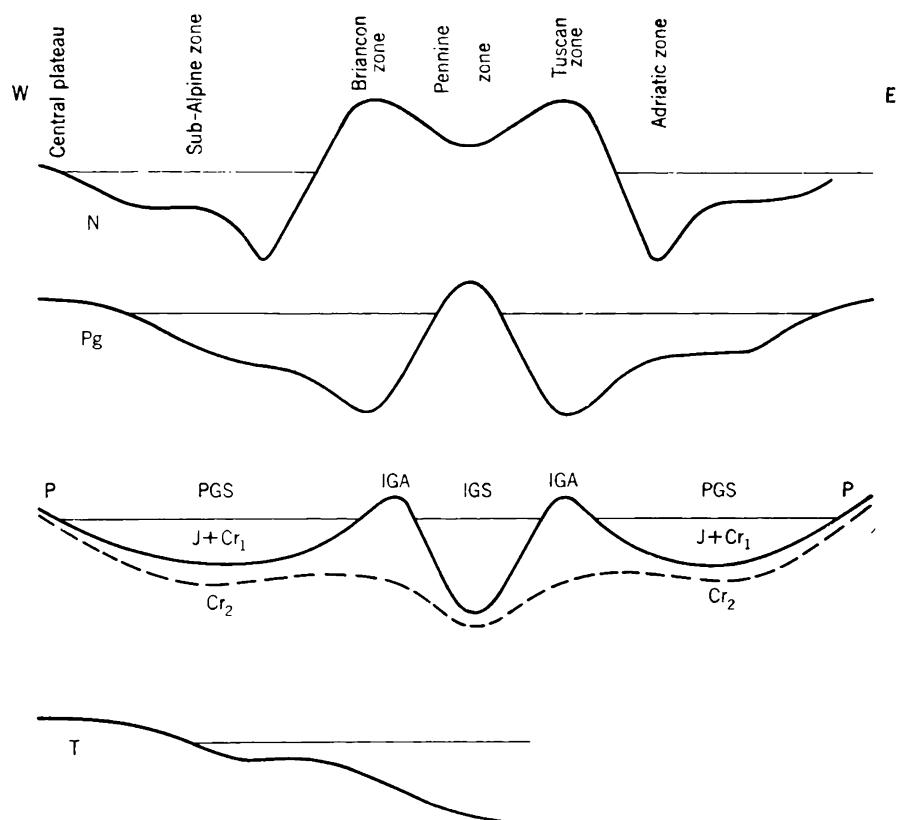


FIG. 306. Development of the western Alps geosyncline and its section: (1) Pre-cambrian; (2) Carboniferous and Permian; (3) Triassic; (4) Jurassic; (5) Cretaceous; (6) Eocene; (7) Oligocene; (8) tectonic faults.

complicated by a brief and intensive uplift of the margin of the central plateau and the Briançon intrageanticline; for this reason no basal Upper Cretaceous sediments were deposited over the Briançon zone. During the second half of the Late Cretaceous (Late Cenomanian), however, very thick carbonate sediments were deposited there comparable to deposits in the intrageosynclines. Therefore, by Late Cenomanian time the intrageanticline must have subsided, in connection with the expansion and rejuvenation of the adjacent intrageosynclines. Great changes occurred at the beginning of the Eocene, when the Briançon zone became the site of a very thick (more than 1,500 m) accumulation of flysch. The sediments become coarser toward the east: to judge by the composition of the pebbles, the clastic material came from the Pennine zone, which must accordingly have been undergoing uplift and erosion. Eocene sediments are absent over the Pennine zone. The subsidence of the subalpine zone continued but less intensively than in the Briançon zone.

Richter (1939) wrote:

Thus the Laramide epeirogenic disturbance brought about a complete reversal of the movements: an uplift within the pre-existing orthogeosyncline (the Pennine), and an intensive subsidence of the Briançon swell. Evidently these processes led to the formation of the Subpennine foredeep, in connection with the pre-Tertiary orogeny in the inner Alps.

Although he uses a terminology somewhat different from ours, Richter clearly has in mind the phenomenon called *inversion* in this book.

Very similar events were occurring simultaneously on the opposite flank of the geosyncline. The Upper Cretaceous of the Pennine zone is represented by limestones, up to 750 m thick. In the Tuscan zone its thickness does not exceed 200 m; nevertheless, the Eocene flysch of the Tuscan zone reaches 3,000 m in thickness, and the clastic material comes from the west, from the Pennine zone.

There is good reason to believe that an inversion took place in this geosyncline toward the end of the Late Cretaceous: its intensively downwarped central part underwent uplift, and the preinversion intrageanticlines became zones of subsidence.

The further development of the geosyncline took the form of expansion of the central uplift, displacing the adjacent basins to the side and causing them to migrate toward the periphery of the geosyncline. In the west, this expansion leads to uplift of the Briançon zone during the Oligocene, whereas subsidence took place over the area of the subalpine parageosyncline, with an accompanying deposition of thick molasse sediments. In a similar development in the east, the expanding front of the central uplift displaced the subsidence from the Tuscan into the Lombard zone and finally into the area of the present Adriatic Sea.

At the present time, the preinversion (Mesozoic) complex of sediments of the axial part of the geosyncline, in the Pennine zone, is unconformably overlain by Oligocene deposits. The bottom of the Oligocene descends stratigraphically toward the center of the zone (around Genoa) and rises to the east and west, toward the Tuscan and the Briançon zones, where Oligocene beds are lacking. They reappear on both margins of the geosyncline, in the younger foredeeps. The distribution and position of the Oligocene beds suggest that after a great terminal Cretaceous uplift in the Pennine zone, a subsidence took place there at the beginning of the Oligocene. The subsidence continued so that the Oligocene sediments were downwarped after their deposition.

Let us review the development of the Alpine geosyncline throughout the Western Mediterranean area. The Pennine intrageosyncline apparently extends into the area of the Tyrrhenian sea, as indicated by the increase in the thickness of the Mesozoic west of the Apennines (Behrmann, 1936). After the inversion, a central uplift appeared in that area; a source of clastic material in the west is indicated by the facies change in the thick Eocene flysch, accumulated during the subsidence that began then, in the area of the present Apennines. This uplift expanded until, in the Oligocene, its eastern margin embraced the zone of the present Apennines. In the Miocene (more precisely, in the Sarmatian age), a subsidence took place within the expanding central uplift. This was of the same nature as that in the Genoa area of the Pennine zone but of greater amplitude, allowing the sea to enter. Consequently, as it stands now, the Adriatic Sea represents the foredeep, the Apennines a part of the central uplift, and the Tyrrhenian Sea an interior depression. Other parts of the central uplift are Corsica and perhaps Sardinia.

Not only the Tyrrhenian Sea but the entire zone between the Balearic Islands and the southern coast of Spain, on one side, and Africa and Sicily, on the other, is bordered by mountain ranges, as if it were a basin in their midst. These ranges are the Apennines, the Sicilian Mountains, the Atlas Range, the Rif Mountains, the Baetic Cordillera, the Balearic Islands, northern Sardinia, and Corsica. All these highlands are marginal uplifts surrounding an interior depression. Beyond them one may distinguish the foredeep, extending from the Adriatic through the southern tip of Italy, along the southern coast of Sicily and thence along the Atlas foothills (Anderson, 1936; Solignac, 1929-1930). It is also encountered along the northern slopes of the Baetic Cordillera and the valley of the Guadalquivir River, whence it descends into the sea north of the Balearic Islands (Stille, 1934). On the dry land the foredeep is everywhere represented by a zone of thick Oligocene and younger Tertiary deposits.

The history of the Western Mediterranean thus appears to be as follows: the Pennine intrageosyncline extended from north to south, across

the Tyrrhenian Sea, turning west between Sardinia and Sicily and coming to a blind end at the Straits of Gibraltar. A preinversion subsidence took place, involving the adjacent portions of what are now the continents. Toward the beginning of the Paleogene an inversion began. Very intensive marginal deeps were formed along the periphery, where the above-mentioned marginal uplifts are presently located. These foredeeps were filled chiefly with Paleogene flysch deposits. Later, with the expansion of the central uplift, all these areas were drawn into it, and the subsidence was displaced outward to its present location. In the Miocene the central part of this great uplift subsided and was invaded by the sea. This latest subsidence, with the formation of an interior depression, extended over various parts of the geosyncline, in places even beyond its limits, and became a more general feature than the local uplifts.

The history and structure of the Celtiberian part of the Alpine geosyncline have been discussed in a previous chapter.

Three areas of subsidence can be distinguished in the Mesozoic history of the eastern Alps: the northern, the central, and the southern limestone areas (Belousov, Gzovskiy, and Goryachev, 1951). In the central depression, a typical intrageosyncline, the Mesozoic sediments are shales, bedded limestones, and dolomites. The inversion occurred here before the Late Cretaceous. The northern and the southern limestone basins contain chiefly Triassic limestones of great thickness, along with Jurassic and Early Cretaceous marls, limestones, sands, and clays. The Upper Cretaceous is represented chiefly by terrigenous deposits originating in the erosion of the uplift which had appeared in the central zone.

Simultaneously with the uplift in the central zone in the Late Cretaceous, and particularly at the beginning of the Tertiary, when the limestone zones were uplifted, new marginal basins were formed symmetrically on the northern and southern edges of the eastern Alps, where an Upper Cretaceous and Eocene flysch of great thickness was deposited, consisting of interbedded sandstones, marls, and shales. Still later, in the Oligocene, when the flysch basins were involved in the expanding uplift of the eastern Alps, new foredeeps developed along its periphery; these, in turn, were filled with thick Oligocene and Miocene molasse deposits.

Between the three limestone zones there were two intrageanticlines during the Mesozoic: the so-called northern and southern graywacke zones, with their much thinner Mesozoic sections and extensive outcrops of older Paleozoic rocks dislocated during the Hercynian and Caledonian cycles.

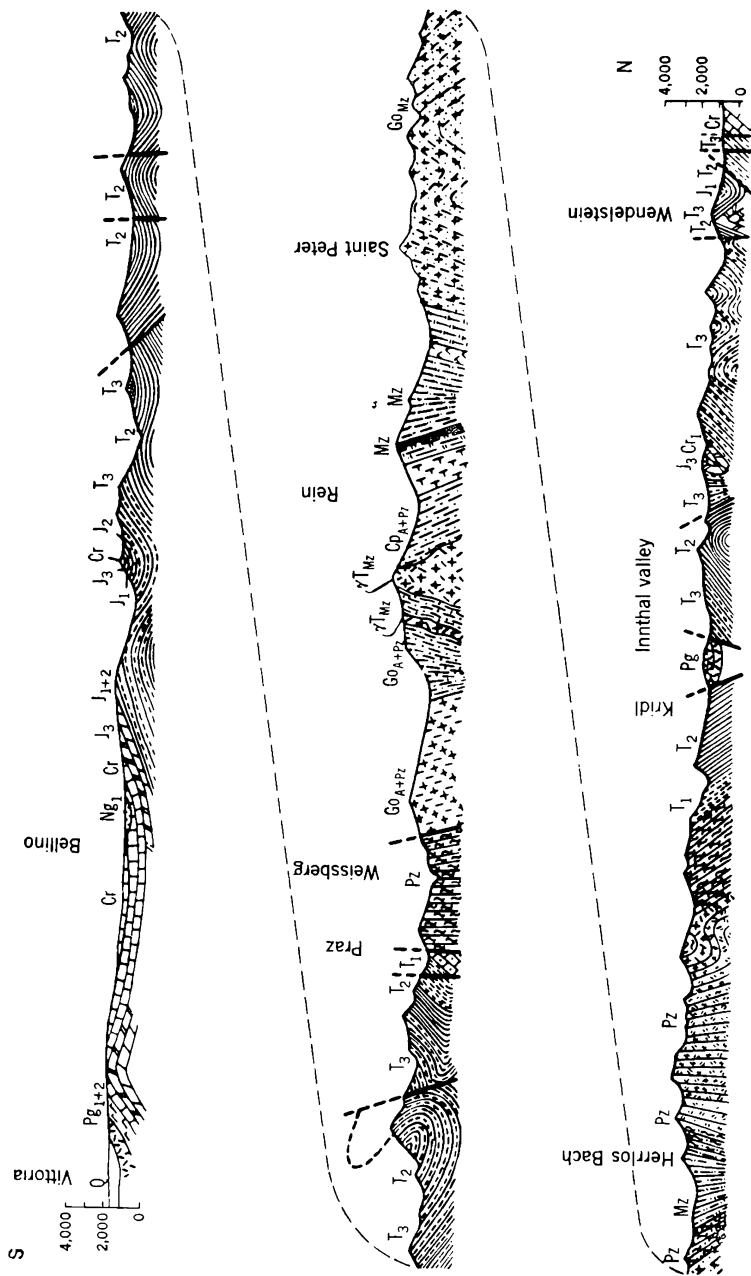
It has been believed for a long time that the structure of both the western and the eastern Alps is characterized by an extensive develop-

ment of very large tectonic thrust sheets, or nappes. Their horizontal displacement in the western Alps was believed to be many tens of kilometers, whereas in the eastern Alps, according to L. Kober, the displacement reached 200 km. It was observed at the beginning of this book, however, that this excessive interest in nappes turned out to be erroneous, based as it was on a fallacious method of drawing geologic conclusions. The current conception of Alpine structure is considerably less esoteric, although local horizontal displacements up to 15 to 20 km are still acknowledged. The distribution of the folds in the Alps reveals the same differentiation of the geosyncline into interior zones of subsidence and uplift: the folds are overturned away from the intrageosynclines and toward the intrageanticlines, and the same is true of the direction of movement of the thrusts. Thus Suess's idea of a single northward "movement of masses" in the Alps is completely erroneous. Actually there are several axes along which the folds are inclined away toward both sides.

A large role in the structure of the Eastern Alps was played by younger normal faults formed chiefly during the Late Tertiary, in the uplift of the mountains. The resulting grabens are sites of some present-day valleys and basins of thick clastic and coal-bearing deposits (Fig. 307).

In the east, the eastern Alps plunge toward the Hungarian basin; only their northernmost and southernmost tectonic zones continue eastward, as the Carpathians and the Dinaric Alps, respectively, encircling the Hungarian basin. This whole area of the Hungarian oval is insufficiently known, and a detailed discussion is impossible. Roughly speaking, its development during the Alpine cycle was as follows (Muratov, 1949; Gzovskiy, 1951): a great geanticline preceded the Hungarian depression during the Mesozoic; it was characterized by thin sediments and by an incomplete section. At the same time in the Carpathians and the Dinaric Alps and over the Balkan Peninsula there were zones of intensive subsidence and thick deposits. During the Mesozoic, these zones expanded gradually; one of them, on the site of the present-day Carpathians, bordered on the Russian platform, where there was a small accumulation of sediments.

In the Tertiary, especially from the Miocene on, the movements were generally inverted. The Hungarian intrageanticline underwent intensive subsidence and became a large intermontane basin with a thick accumulation of Paleogene, Miocene, and especially Pliocene sediments, partly marine and partly lacustrine and terrestrial. Simultaneously, the subsidence in the Carpathians, the Dinaric Alps, and the Balkan Peninsula was replaced by uplifts that, toward the end of the Tertiary, resulted in the formation of mountain ranges bordered by foredeeps. The foredeep along the outer arch of the Carpathians, where it, as



usual, draws the margin of the platform into its subsidence, is very readily discernible. Typical molasse sandy-argillaceous and lagoonal evaporite formations accumulated in this foredeep. In the course of the uplift, the sedimentary strata of the Carpathians, the Dinaric Alps, and the Balkan Peninsula were strongly folded, the predominant direction of overturning and thrusting being toward the platform. Along the inner side of the newly risen mountains, overturned toward the Hungarian intermontane basin, many faults were formed, allowing escape of great volcanic flows.

This very simple scheme is actually considerably complicated by secondary and smaller structures, of which only a few will be mentioned. The extensive Hungarian Mesozoic intrageanticline is not homogeneous. Within it lies a zone of comparatively strong subsidence and thick

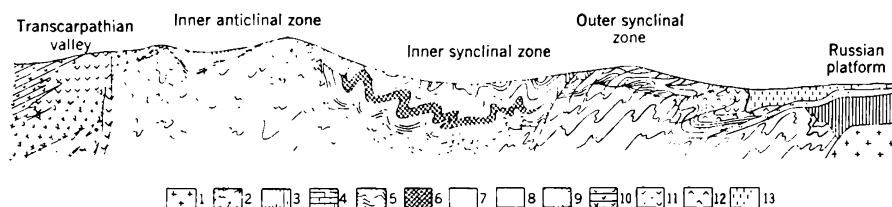


FIG. 308. Structural section through the eastern Carpathians (after M. V. Muratov): (1) Precambrian basement of the Russian platform; (2) Precambrian and Paleozoic rocks of the Carpathian basement; (3) Paleozoic rocks of the platform; (4) Jurassic; (5) Lower Cretaceous; (6) lower strata of the Upper Cretaceous; (7) Senonian (flysch); (8) Upper Cretaceous platform facies; (9) Paleogene; (10) Miocene and Pliocene of Transcarpathia; (11) effusives of Transcarpathia; (12) Lower Miocene of Ciscarpathia; (13) Tortonian and Sarmatian of Ciscarpathia.

Mesozoic sediments (Triassic, up to 2,000 m; Jurassic, 250 m; Cretaceous, 330 m) where there was subsequently an uplift, accompanied by folding of the discontinuous or intermediate type and by the formation of the upland known as the Central Hungarian Mountains. The latter have divided the huge intermontane basin into the Greater and Lesser Hungarian basins, whose evolutionary histories differ somewhat.

A similar local Mesozoic subsidence took place in the area of the present Mecsek Mountains (south of Lake Balaton). Such local areas of subsidence can be considered parageosynclines of an order below that of the Hungarian intrageanticline.

The Greater Hungarian basin began its subsidence in the Paleogene and the Lesser Hungarian basin generally in the Pliocene. In both, the point of maximum subsidence and greatest accumulation migrated with time during the Pliocene, thus causing a lack of correspondence in areas between the several local stratigraphic sections (Gzovskiy, 1948).

The most thoroughly studied is the part of the Carpathians within the U.S.S.R. (Fig. 308). According to A. A. Bogdanov (1949b) and M. V.

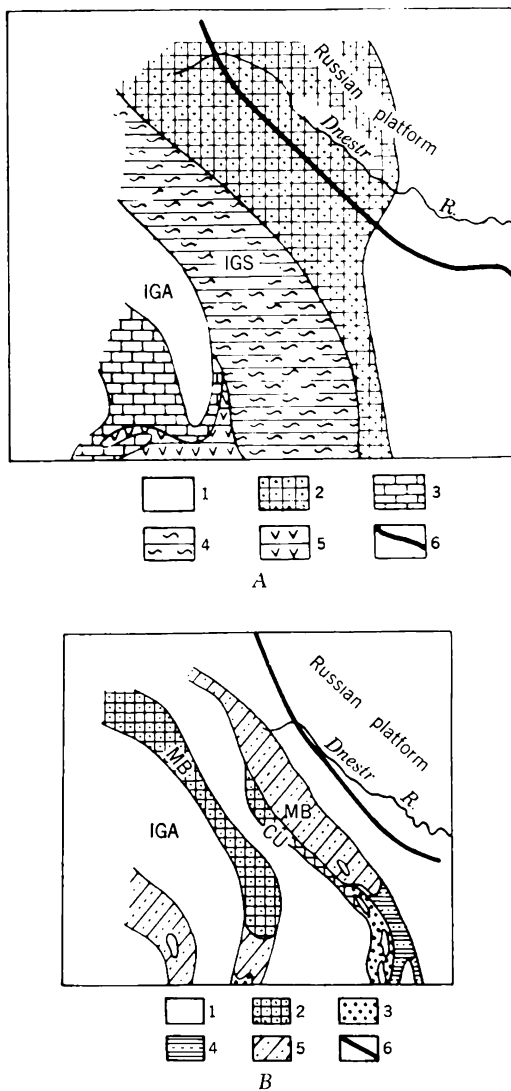


FIG. 309. Diagrams showing the history of the eastern Carpathians (*after M. V. Muratov*). (A) Lusitanian structure of the Carpathians: (1) areas of erosion; (2) shallow-water limestones; (3) thick shallow-water-limestone series; (4) silicate jasper rocks alternating with limestones; (5) effusives; (6) limit of Precambrian basement of Russian platform. (B) Structure of the eastern Carpathians in the Aptian: (1) areas of erosion; (2) shallow-water limestones; (3) conglomerates and sandstones; (4) sandy-argillaceous formations; (5) sandy-argillaceous flysch; (6) limit of Russian platform. (C) Eastern Carpathian structure in the Middle Eocene: (1) areas of erosion; (2) littoral sands and sandstones; (3) shallow-water limestones; (4) conglomerates and sandstones; (5) thick sandstones; (6) sandy-argillaceous flysch; (7) limit of Russian platform. (D) Eastern Carpathian structure in the Tortonian: (1) areas of erosion; (2) mountains; (3) swamp and lacustrine deposits; (4) active volcanos above sea level; (5) shallow-water limestones; (6) clays with sand interbeds; (7) reef limestones; (8) sandy and sandy-argillaceous deposits; (9) extrusive tuff and lava rocks; (10) limit of Russian platform.

Muratov (1949). the history and the structure of this portion of the Carpathians are as follows: the Alpine subsidence of the earth's crust in the eastern Alps began in the Triassic, when the strongly dislocated Paleozoic beds were overlain by multicolored sandstones and conglomerates, followed by dolomites, marls, and siliceous and sandy-argillaceous sediments. The subsidence occupied the central and axial parts of the Carpathians; along their margin, in Ciscarpathia, no sediments were deposited (Fig. 309). These conditions persisted into the Jurassic, when the subsidence continued in approximately the same zone, although it extended to the east, involving the edge of the platform. This expanding intrageosyncline was filled chiefly with carbonate deposits and, to less

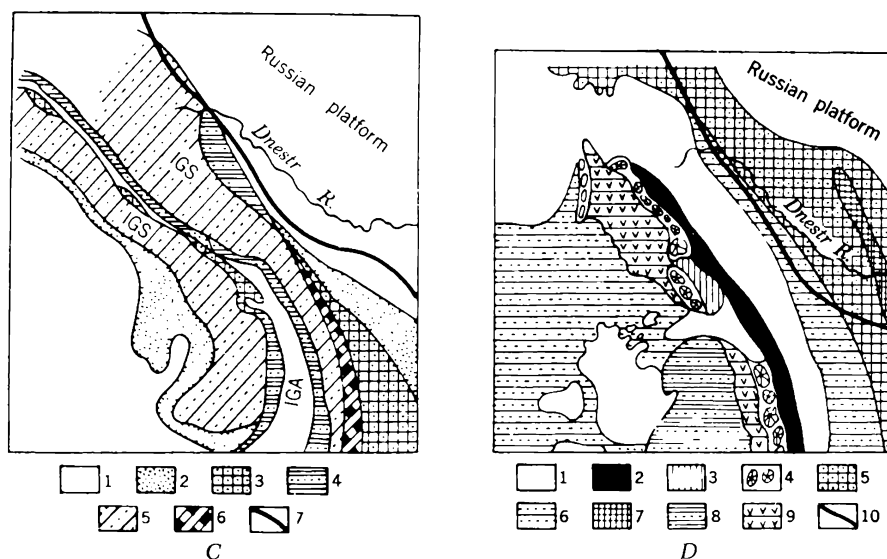


FIG. 309 (Continued)

degree, with argillaceous and sandy sediments, all cut through by diabases.

A change in conditions marked the beginning of the Lower Cretaceous, when a narrow uplift appeared within the area of subsidence and divided the intrageosyncline into two parts. This uplifted zone follows the trend of the present Marmarosa massif, forming the interior anticlinorium of the eastern Carpathians. Intensive subsidence continued on both sides of this central uplift, with the accumulation of a very thick flysch consisting of alternating sand and clay beds, along with marls and limestones. The well-exposed eastern zone of flysch deposits, which in the present structure of the eastern Carpathians corresponds to the so-called middle synclinorium, has been extensively studied. Over a long distance on the western (inner) slopes of the

Carpathians, the flysch is dropped down along large faults and covered by younger rocks; it crops out in this zone outside the Soviet Union, in Romania. This interior depression differs from the outer one in its smaller size and depth of subsidence.

The above conditions, except for local uplifts and subsidences, persisted throughout the Cretaceous and Paleogene. The situation changed once more at the end of the Paleogene and the beginning of the Neogene. At that time, the eastern flysch zone of subsidence was uplifted and received no further sediments. But to the east of it a new zone of intensive subsidence appeared. This was a typical foredeep, in which Miocene and Oligocene sandy-argillaceous and salt-evaporite deposits accumulated. West of the central uplift, along the interior slope of the Carpathians, the subsidence also was intensified, accompanied by the accumulation of sediments typical of foredeeps and intermontane basins: salt-evaporite and molasse deposits as well as extensive volcanic flows. This western Neogene basin was located in the approximate area of an earlier flysch basin, and its axis was not perceptibly displaced during the Neogene. The subsidence occurred to a considerable extent along deep high-angle faults, which apparently served as channels for lavas and acidic intrusions. The eastern (outer) foredeep gradually migrated throughout the Neogene toward the platform, as if it were being pushed aside by the constantly expanding Carpathian uplift.

The present structure of the eastern Carpathians has the following major elements: an interior (marginal) basin or synclinorium; an inner anticlinorium corresponding to the central uplift appearing at the beginning of the Cretaceous; an outer anticlinorium corresponding to the uplift appearing at the margin of the flysch zone at the close of the Paleogene-Neogene; and an outer foredeep or synclinorium. Against the background of these major structural elements, there was intensive folding in many phases during the Cretaceous, Paleogene, and Neogene. Sharp asymmetry characterizes the entire eastern Carpathian uplift: east of the axis of the inner anticlinorium, the direction of movement as revealed in the overturning of the folds is clearly to the east; in the outer anticlinorium, great overthrusts of several kilometers are displaced in the same direction. At the same time the westward inclination of the folds, which one would expect to find along the inner flank of the Carpathians, is very slight.

Thus the general succession noted elsewhere is also observable in the development of the eastern Carpathians: a predominance of subsidence in the beginning of the cycle, a local inversion with the formation of a central uplift and its subsequent expansion, the development of foredeeps, and a general uplift at the close of the cycle. The expansion of the central uplift, as in several previous instances, is sharply asymmetrical—its direction is almost exclusively to the east, with almost no

expansion westward. The change in the facies is also regular, from a very thin lower terrigenous sequence (Lower Triassic) to limestone (Triassic-Jurassic) in the first half of the cycle and from upper terrigenous (Cretaceous and Paleogene flysch) to lagoonal and molasse (salt-evaporite and sandy-argillaceous Neogene formations) in the second half.

Intermediate and discontinuous folding is widely developed along the outer margins of foredeeps, with numerous diapir folds. In Czechoslovakia and Romania the eastern Carpathians have been studied in less

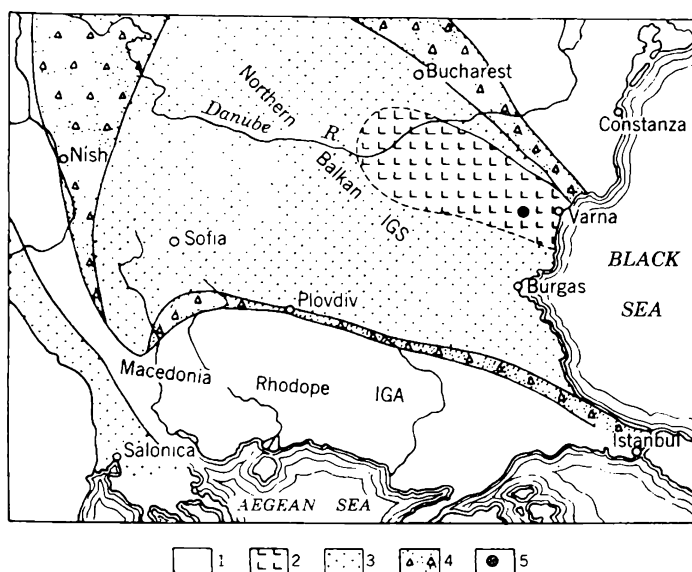


FIG. 310A. History of the structure of the Balkans (*after* M. V. Muratov). The Balkans in the Late Permian and Early and Middle Triassic: (1) areas of erosion; (2) salt-evaporite deposits of the Late Permian and Early Triassic and Middle Triassic limestones; (3) Late Permian and Early Triassic clastics and Middle Triassic limestones; (5) salt dome.

detail than in the U.S.S.R. Proper understanding of their history and structure is hindered by the acceptance of the nappe theory. Judging from the most general data, the overall development is similar throughout the Carpathians.

The mountain ranges of the northeastern Balkans are a continuation of the Carpathians. Data on this region have been compiled by M. V. Muratov (1949). On the south and west this area borders on the Macedonia-Rhodope massif of crystalline and Paleozoic rocks. Within the framework of the geosyncline, at the beginning of the Alpine cycle, this massif became an intrageanticline—the direct southeastward extension of the Hungarian intrageanticline (Fig. 310). This intrageanti-

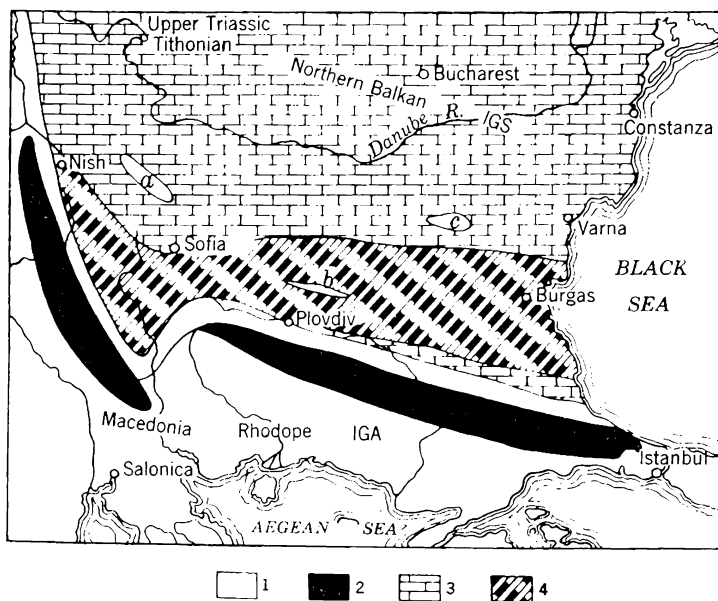


FIG. 310B. The Balkans in the Tithonian: (1) areas of erosion; (2) mountains; (3) thick shallow-water limestones; (4) carbonate flysch; (a, b, and c) isolated uplifts.

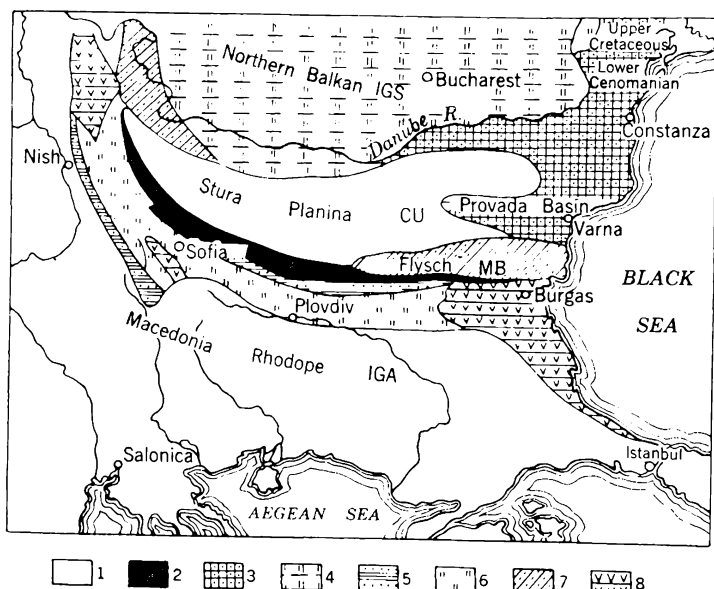


FIG. 310C. The Balkans in the Early Cenomanian: (1) areas of erosion; (2) mountains; (3) shallow-water limestones; (4) deep-water marl-limestone formations; (5) thick sandstones; (6) thick marls, limestones, and sandstone; (7) flysch; (8) extrusives.

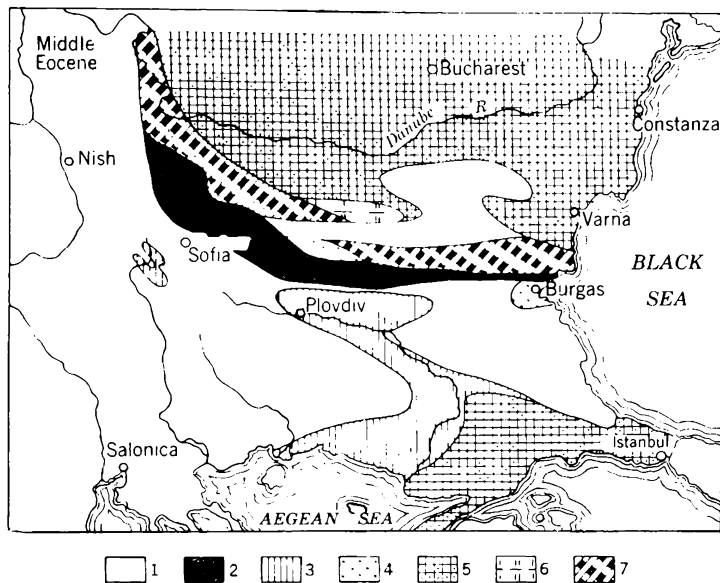


FIG. 310D. The Balkans in the Middle Eocene: (1) areas of erosion; (2) mountains; (3) coal-bearing continental deposits; (4) littoral sands; (5) thin shallow-water limestones; (6) limestones and marls; (7) flysch.

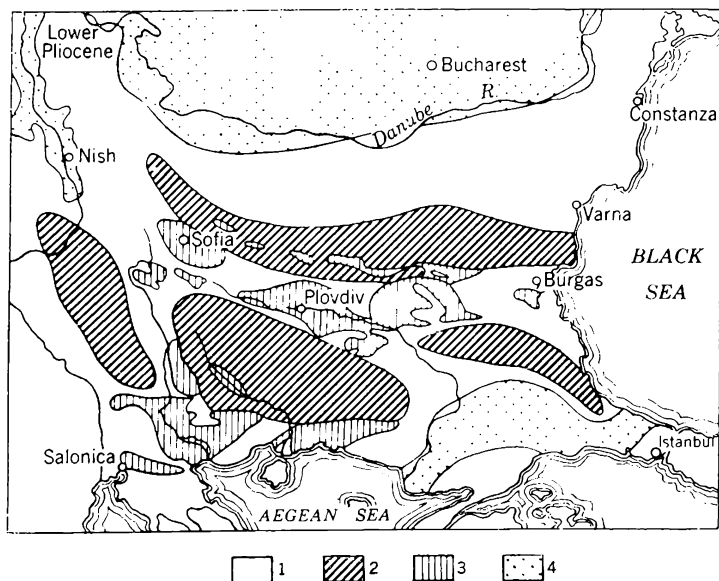


FIG. 310E. The Balkans in the Pontian: (1) areas of erosion; (2) mountains; (3) lacustrine and fluvial deposits, with coal; (4) sands and sandstones.

cline was apparently formed at the end of the Paleozoic, as a central uplift in the Hercynian cycle. To the east it continues into Turkey and merges directly with the intrageanticline of the Taurus-Caucasus region, which has been named the *Araks intrageanticline*. Its continuation can be traced in the interior of Iran. Thus this intrageanticline is an important element in the structure of the Alpine geosyncline of south-eastern Europe and the Near East, forming what A. D. Arkhangel'skiy has called a *central massif*.

Both north and south of the Macedonia-Rhodope intrageanticline are areas of subsidence, the Balkan intrageosynclines, which formed as intermontane basins at the end of the Hercynian cycle. Deposition of the Alpine cycle began in the northern Balkan intrageosyncline with an accumulation of red sandstones and salt deposits in the Late Permian and Early Triassic. These deposits may be considered as the terminal (molasse and lagoonal) foredeep deposits of the previous cycle. They were followed by continental deposits, partly coal-bearing, and then by sandstones, limestones, and slates in the Early and Middle Jurassic, attaining great thickness in this basin. In the Tithonian an intensive uplift of the Macedonia-Rhodope intrageanticline took place, and the axis of maximum subsidence of the northern Balkan intrageosyncline was displaced toward the foot of the uplift. Here a great thickness of carbonate flysch began to form, consisting of alternating marls, limestones, and clastic sediments. To the north, the flysch passes into thinner shallow-water limestones. Against this background, there were additional multiple fluctuations of the zones of uplift and subsidence. At the end of the Cretaceous the entire region underwent a temporary general uplift.

Considerable changes occurred, beginning in the Late Cretaceous, when a new uplift—the Stara Planina uplift—appeared in the axial part of the northern Balkan intrageosyncline, and the subsidence continued both to the north and south. There can be no doubt that a local inversion took place here, within the northern Balkan intrageosyncline, resulting in a central uplift and two foredeeps on its flanks: the Danubian and the Kraishtida-Burgas basins (Fig. 310C). The first originated along the outer margin of the geosyncline, on the edge of the platform, whereas the second occurred on the northern flank of the Macedonia-Rhodope intrageanticline. The Upper Cretaceous deposits are highly diversified, represented by sandstones, clays, marls, and limestones and partially by a carbonate flysch. Throughout the Late Cretaceous the Stara Planina central uplift was intensively elevated and eroded. During the Eocene a thick flysch was deposited on both sides of the central uplift, under similar geotectonic conditions.

During the Neogene, the Stara Planina central uplift continued to expand, so that the adjacent zones of subsidence were displaced out-

ward. The northern basin was displaced into the depression of Wallachia, where a typical foredeep was formed, with its characteristic molasse deposits. Similarly the southern, or Kraishtida-Burgas, subsidence was displaced southward, "rolling" over the Macedonia-Rhodope intrageanticline. In this process it was divided into a number of individual, isolated basins, producing the effect of separate depressions over the surface of the intrageanticline. This is an example of the process by which an intrageanticline becomes an intermontane basin—a process that is not fully realized here but merely indicated by a series of separate depressions. In the depressions adjacent to the Sea of Marmora and the Aegean Sea, Eocene and Neogene marine deposits accumulated. These were mostly sand and clay and were locally very thick. The depressions isolated in areas more remote from the sea were filled with lacustrine and alluvial sediments. Continental deposits predominate in the Neogene and particularly the Pliocene, both in the Wallachian foredeep and in the series of intermontane basins over the Macedonia-Rhodope massif. A general uplift of the area took place at the end of the Tertiary and beginning of the Quaternary.

In the present structure of this region, the Macedonia-Rhodope massif is, generally speaking, a noninverted anticlinorium complicated by individual inverted synclinoria. The Stara Planina region, corresponding to the central uplift, is reflected as an inverted anticlinorium, and the Wallachian has become a synclinorium formed on the site of a foredeep. Intensive but little known folding is developed in the Stara Planina area; to the north, this gives way to the intermediate and discontinuous folds of the Wallachian depression. The Alpine folding within the basins scattered over the Macedonia-Rhodope massif is also not intensive and is represented chiefly by domical uplifts of the discontinuous type. This manifestation of an already familiar sequence in the development of geosynclines is so obvious that no additional comments are necessary.

The area west and south of the Macedonia-Rhodope massif is too little known for an adequate analysis of its evolution. One of its elements is apparently the intrageosyncline of the Dinaric Alps, situated symmetrically in relation to the northern Balkan intrageosyncline and reflecting a similar development. In the directions of the Adriatic Sea and the Peloponnesus, the continuous folds are replaced by intermediate and discontinuous folds, suggesting the proximity of a platform. The Adriatic Sea is a foredeep in relation to both the Apennines and the Dinaric Alps.

The Alpine geosyncline includes the Crimea. The most complete analysis of the evolution and structure of this peninsula (Fig. 311) has been presented by M. V. Muratov (1949). Toward the end of the Hercynian cycle, an uplift began north of the present Crimean moun-

tains, in the steppe region of the Crimea. According to M. V. Muratov, this uplift was connected with the Dobruja to the west. South of this area, on the site of the present Crimean mountains, there was a subsidence that developed into an intrageosyncline of the Alpine cycle. Deposition in the Crimean Alpine intrageosyncline began in the Triassic

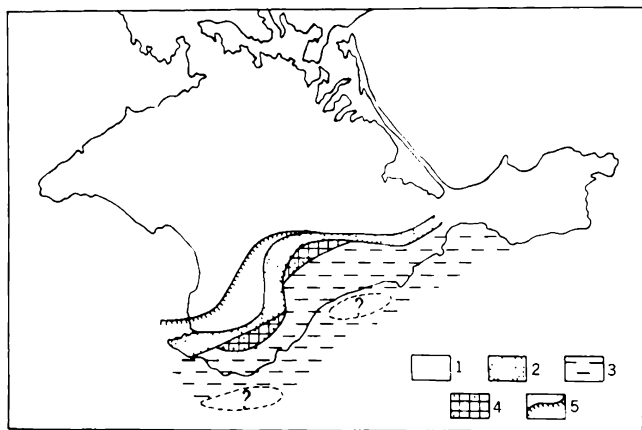


FIG. 311A. The Crimea in the Valanginian age: (1) areas of erosion; (2) calcareous sands and sandstones, with interbedded limestones and marls; (3) clays and shales; (4) shallow-water sandy limestones; (5) inferred limit of Hauterivian deposits.

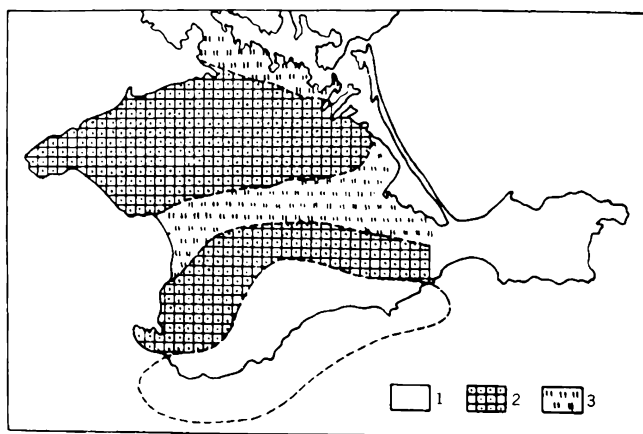


FIG. 311B. The Crimea in the Cenomanian: (1) areas of erosion; (2) calcareous sands and sandstones with interbeds of clays and marls; (3) deep-water limestones and marls.

with the thick "Taurian formation" represented by interbedded shales and sandstones. The sandy-argillaceous sediments of the Lower and Middle Jurassic and Lower Cretaceous systems were deposited above these. Throughout this time the present steppe area of the Crimea, north of the Crimean intrageosyncline, was occupied by an uplifted

massif: the southern margin of the Russian platform, which was being eroded.

Toward the very end of the Early Cretaceous, in the Albian and very definitely in the Late Cretaceous, the conditions changed: in the Crimean intrageosyncline, a central uplift was formed in which the development was sporadic. Uplift was comparatively weak during the Late Cretaceous and Paleogene, when clastic sediments were deposited only in its immediate vicinity, whereas marls and limestones predominated only a short distance to the north. Several times during the Late Cretaceous the Crimean uplift was fully submerged. It was strongly revived in the second half of the Tertiary, however, when its erosion

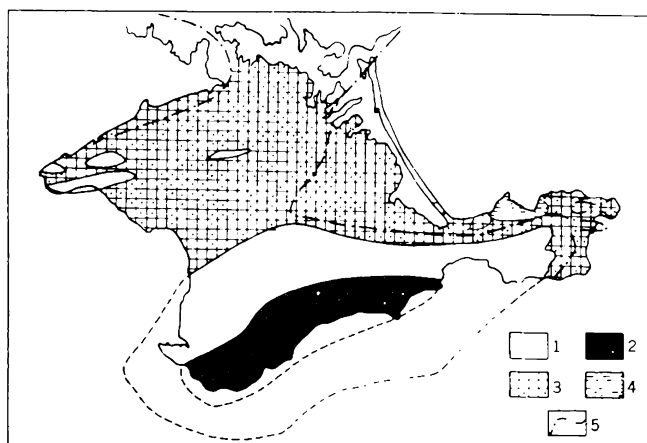


FIG. 311C. The Crimea in the Meotian: (1) areas of erosion; (2) uplands; (3) shallow-water limestones; (4) clays; (5) inferred shore line at the beginning of the Meotian.

supplied a vast amount of clastic material; a general uplift of the present Crimean Peninsula took place in the Quaternary.

The axis of the central uplift, which had appeared in the beginning of the Late Cretaceous and continued to develop in the Tertiary, was located south of the present southern shore of the Crimea. Apparently the uplift was asymmetrical: its southern flank was considerably steeper than its northern.

This intensive uplift led to the formation of a fault in its southern flank, a high-angle thrust that is still present as the locus of a zone of earthquake foci (see Fig. 1). In the second half of the Pliocene, subsidence of the uplifted arch began, accompanied by marine erosion on the south. Thus only the northern flank of the uplift is now on dry land.

The central uplift of the Crimean Mountains is an anticlinorium. The shales of the Taurus suite are intensively and disharmonically folded. The younger deposits are considerably less disturbed, and their position suggests that the anticlinorium was only slightly complicated by a few

gentle secondary anticlines and synclines, which at least in the eastern Crimea have the character of injection folds. It is also broken by a number of normal faults, whose strikes are mainly perpendicular to the trend of the folds.

Igneous activity in the Crimea is represented by basic flows from the beginning of the cycle, in the Triassic and Middle Jurassic, as well as by minor intrusives of alkaline composition, whose history apparently is closely connected with the uplift of the anticlinorium.

North of the southern European Alpine geosyncline lies the Alpine platform (see Arkhangel'skiy, 1947-1948), which overlies a variegated folded basement. Over a large area, including the Russian plain and the countries surrounding the Baltic Sea, the Alpine platform rests on a Precambrian folded basement. In Norway, throughout nearly all of England, and in northwestern Germany, this basement is of Caledonian age; over the remainder of western and southwestern Europe, it is Hercynian. A Hercynian basement also underlies the Donets basin.

This enormous Alpine platform is broken up into a large number of synclises and antecises, which, from the historical standpoint, are subgeosynclines and subgeanticlines, respectively. In the north, the large Fennoscandian antecise is divided by the Precambrian crystalline rocks that crop out in its center. In the east, trending north-south, lies the eastern Russian syncline, of which the Moscow and Dnepr-Donets synclises are branches. The latter is separated from the Polish-German syncline by the Poleska antecise, which, in turn, is a southern branch of the Fennoscandian antecise. Farther to the south and southwest, the Subhercynian, Thuringian, northwest German, Parisian, and Aquitanian synclises have been distinguished, with the Bohemian, Hartz, Rhine, Armorican, central plateau of France, and Spanish Meseta antecises, respectively, between them. In England, the Hampshire and London synclises are recognized; the remainder of the country generally represents a large antecise, with outcrops of the Caledonian folded basement.

Not all these synclises and antecises are of the same age. Those overlying a Precambrian basement were formed earlier, either in the Caledonian or Hercynian cycle; those overlying a Hercynian basement originated at the beginning of the Alpine cycle. As a rule, the structural expression of the more ancient synclises and antecises is weaker than that of the younger ones, suggesting that in the course of time the differential oscillatory movements over the platform are damped down and diffused.

In the synclises there was an accumulation of Mesozoic and Cenozoic sediments, as a result of their subgeosynclinal regime of movements; the greatest thicknesses occurred in the areas of maximal subsidence. When a subgeosyncline persisted through several cycles (three, in the case of

the Precambrian basement), Paleozoic formations that accumulated under similar subgeosynclinal conditions are found under the Mesozoic sediments. In applying this rule, however, one must take into account the migration of the subgeosynclines, or rather of the zone of their greatest subsidence, as mentioned previously in discussing the general rules of platform development. We have seen, for instance, that the Baltic Caledonian subgeosyncline had moved southward by the beginning of the Hercynian cycle to become the Moscow subgeosyncline. Hence the zones of maximal deposition in the Early Paleozoic, on the one hand, and in the Middle and Late Paleozoic, on the other, do not coincide in that part of the platform. Displacements of the zones of greatest subsidence from epoch to epoch are observed in other subgeosynclines as well. Thus the present structure expressed in the existing synclises and anteklises is the net result of all the previous movements and, as such, may be somewhat different at different levels. Generally, however, such differences are not likely to be great.

Synclises contain discontinuous folds such as swells and domes. Locally, as in southern Germany, this folding is very intensive (*Saxon folding*) and has the character of box folds.

As a rule, the folded basement is exposed in the center of an anteklise. The Alpine structure of the anteklises is manifested to a considerable degree as disruptive dislocations: bending fractures and normal faults.

Intensive jointing is observed in Fennoscandia. The faults complicating the structure of the Rhine anteklise have already been mentioned. Somewhat weaker but still numerous dislocations fracture the central plateau of France. Younger volcanic activity is associated with the faults; this is especially noticeable in southern Germany, Bohemia, and France.

Since the folded basement complex is exposed in the centers of the Alpine anteklises, it might be further differentiated to some degree, wherever it has not been involved in subsequent intensive folding. Such regional subdivision of the Precambrian geosynclines is still impossible, however, and will not be attempted here. The Caledonian folding in Norway is now part of a central uplift, the other part of which must be covered by the waters of the Atlantic Ocean. The structure of the Caledonian geosyncline in England has already been described; three central uplifts have been distinguished there: the Welsh, the southern Scottish, and the northern Scottish uplifts, with two intermontane basins, those of the Lake Region and the central valley of Scotland, between them.

The pattern of Hercynian folding may be restored to a certain extent from the parts exposed in the eroded anteklises of Germany, France, and Spain. Apparently two central uplifts were formed toward the end of the Hercynian cycle. One, in the north, extended from the central plateau through the Vosges and the Black Forest into Bohemia; the

other, in the south, is seen in the Meseta, with a possible continuation into Sicily. The northern central uplift is complicated by small superimposed basins containing limnic coals. North (in Westphalia and Southern England) was a foredeep with a thick accumulation of paralic coal, and an intermontane basin occupied the site of the present Pyrenees.

The position of the Donets basin is of particular interest. Judging from the sediments, the Greater Caucasus was formed on the site of a Hercynian intermontane basin, so that a central uplift must have existed north of the Caucasus, extending into the northern Crimea. If this is true, most of the Donets basin area, according to its position and the character of its sediments, must have been situated within the Hercynian foredeep. But this wedges out toward the northwest, and there are no traces of any geosynclinal development in the Paleozoic deposits of Volhynia. Therefore the Hercynian geosynclinal belt wedges out in the northwest and is not directly connected to the part that lies under the Kielce-Sandomir mountains but rather trends parallel to it en echelon.

ASIA

The structure and history of Asia are of great interest, first, because of the presence of enormous "reactivated" regions with a recent intensification of the differential movements of the earth's crust; and second, because of the widespread intermediate parageosynclinal conditions. But the lack of knowledge of many parts of Asia makes an analysis of its geotectonic structure extremely difficult.

In Asia the Alpine geosyncline embraces Turkey, the Caucasus, Iran, Afghanistan, Korea, the Pamirs, the Himalayas, Indonesia, the Philippines, Japan, Sakhalin, the Verkhoyansk Range, and the region to the northeast of it, including Kamchatka. Moreover, an isolated geosynclinal region encompasses the eastern Transbaykal and the adjoining part of Mongolia.

This very long geosyncline is best known in its Taurus-Caucasus area, the structure and evolution of which have already been discussed. According to the fragmentary data available, the Turkish portion of the Alpine geosyncline is subdivided into a broad central intrageanticline and two intrageosynclines in the north and south. To the south lies the Taurus intrageosyncline, extending westward into the Dinaric Alps of southeastern Europe. The intrageanticline of eastern Turkey is here called the Araks intrageanticline; this is a continuation of the Hungarian and Macedonia-Rhodope intrageanticline; the northern (Anatolian) intrageosyncline passes eastward into the Trialet intrageosyncline.

A similar differentiation exists in Iran and Afghanistan, where both the northern and southern mountain ranges uplifted from the intrageosynclines are counterbalanced by a large central depression repre-

senting a broad intermontane basin, or rather a series of intermontane basins over an intrageosyncline. The general development of this region follows the sequence manifested in the Taurus-Caucasus.

In the Pamirs both the intrageosynclines converge, and the intervening intrageanticline wedges out. Out of this single intrageosyncline rose the Pamirs and the Zaalayskiy Range toward the end of the Mesozoic, whereas the Alayskaya Valley and its continuation, the Tadzhik depression, were a foredeep in that part of the Alpine geosyncline. The folding there is chiefly intermediate, as it is along the slopes into the foredeep, in the Peter I Range. There are reasons to believe that the Alpine geotectonic development in the Pamirs was somewhat less intensive and that it was manifested in parageosynclinal rather than intrageosynclinal conditions. The enormous later uplift of the earth's crust that produced this elevated region is associated with the reactivation embracing not only the southern part of the Alpine platform of central Asia but also part of the area of the Alpine geosyncline. The data for a complete analysis, however, are still lacking.

The eastern continuation of the Pamirs are the mountains of Tibet, the Karakorum, the northern Himalayas, and the Himalayas proper. The structure of this vast mountain region is very little known. In the Himalayas there was a thick accumulation of Mesozoic sediments. In the Tertiary the mountain ranges were uplifted, and a foredeep appeared at their southern foot, containing an accumulation of predominantly Late Tertiary molasse deposits.

The folding in the Himalayas is characterized by a southward overturning of the folds toward the Indian platform. The northward overturning there is slight: it appears only along a short distance in the western part of Karakorum (Wadia, 1926; De Terra, 1936). In Tibet and in the Oman Mountains the inversion apparently took place at the end of the Jurassic or beginning of the Cretaceous.

Still farther on, the Alpine geosyncline extends into the western part of Indochina, where it turns sharply to the south to include the Arakan Yoma and Pegu Yoma ranges. From the continent, the geosyncline reaches to the islands of Indonesia, including Sumatra, Java, the Celebes, and eastern Borneo, and thence to the islands of East Asia, into the Maritime Provinces and northeastern Asia.

The tectonic structure of Indonesia has been described in the well-known works of van Bemmelen (1949), who sees indications of a gradual wavelike outward spread of uplifts from certain centers in the Alpine tectonic history of Indonesia. These centers for the most part were in the area of the present interior seas of Indonesia (the South China Sea, the Japan Sea, the Banda Sea, etc.). The areas of the present islands of Indonesia were occupied mostly by uplifts during the Tertiary; thereafter the centers of this wavelike process, which earlier had been uplifted, subsided and became seas. Sumatra and Java are

composed of uplifted foredeep deposits. The folding in these islands is intermediate and discontinuous; there is extremely violent recent and present-day volcanic activity associated with numerous dislocations of the normal fault type. A present-day foredeep may be seen in the deep Sunda trench, which fringes the outer side of the Indonesian island arc.

The subsidence of the Alpine cycle at the Primorje margin and in the Verkhoyansk region began very early, in the Permian; it continued intensively throughout the Triassic and locally into the Jurassic; toward the Cretaceous, it gave place to uplift.

A study of the tectonic history of Japan shows that at the end of the Paleozoic there was a great uplift of the earth's crust within the Hercynian geosyncline, in the area of the Sea of Japan west of the present Japanese Islands (Kobayashi, 1951). This uplift gradually spread eastward and by the beginning of the Mesozoic had reached the inner side of the Japanese Islands. The outer (eastern) shore of these islands is an Alpine intrageosyncline that was uplifted and folded during the Early Cretaceous. The central uplift in this intrageosyncline was apparently formed in the Barremian; by the Cenomanian and the Turonian there was already an extensive deposition of red molasse. At the end of the Cretaceous, the uplift encompassed all of Japan. It continued into the Paleogene, when only local ingressions of the sea are observed. Throughout the Paleogene Japan was part of the continent of Asia, and the Sea of Japan did not exist; this sea was formed in the Miocene, when Japan also began to be separated into larger and smaller islands. According to G. U. Lindberg (1955), the Sea of Japan is very recent—of Quaternary age. The depression of the present East China Sea, at any rate, did not subside later than the Pliocene. The folding in Japan during the Tertiary was primarily intermediate and discontinuous (box folds, brachyanticlines, and domes).

An extremely interesting problem, which is also of significance for the geotectonics of East and Southeast Asia, is that of the island arcs that fringe the western shores of the Pacific Ocean. There have been many suggestions regarding the causes of their formation, none of which may be considered sufficiently well founded. Because of the scantiness of the available data, this question will not be discussed here; it has already been indicated that the deep ocean trenches that parallel the outer edges of the island arcs are foredeeps that have not yet become filled with sediments.

The considerable area of China that lies outside the geosyncline is distinguished by its peculiar tectonic development (Belyayevskiy, 1949; Sinitsyn, 1948; Chardin and Licent, 1924; Grabau, 1923; Hsieh, 1936*a,b*; Lee, 1939; Leuchs, 1937; T'an, 1936–1937; Ting, 1929; Wong, 1927, 1929; Belousov, 1956; Sinitsyn, 1955).

In central and southern China there is an extensive Alpine platform

whose basement is primarily Archean, although some parts of it have Caledonian (in Tsinan) and Hercynian (in the Nan Shan, Kunlun, and Tsinling areas) folded basements (Fig. 312). The platform is divided into a number of antecises and synecises formed at various times. The history of the oscillatory movements in some of these structural ele-

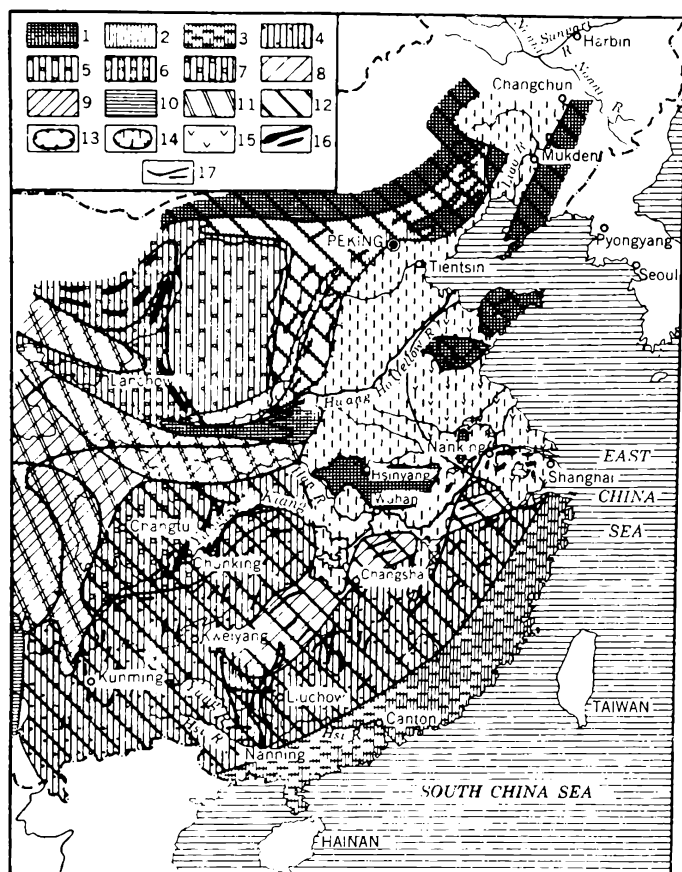


FIG. 312. Diagram of the structure of central and southern China. Platforms overlying Archean basements (1-7): (1) shields; (2) antecises; (3) antecises undergoing granitization in the Cretaceous; (4) synecises considerably subsiding in the Devonian, Carboniferous, and Permian and uplifted in the Cretaceous; (5) synecises undergoing considerable subsidence mainly in the Permian, but also in the Triassic and Cretaceous; (6) Hercynian foredeeps continuing their subsidence in the Mesozoic and Cenozoic; (7) synecises undergoing considerable subsidence in the Triassic, Jurassic, and Cretaceous; (8) antecises resting on a Caledonian basement; (9) antecises resting on a Hercynian basement; (10) Alpine geosyncline with a Mesozoic inversion; (11) regions of intensive tectonic differentiation in the Cretaceous; (12) regions whose tectonic differentiation ended in the Cretaceous. Areas of recent tectonic activity (13-15): (13) areas of considerable uplift; (14) areas of considerable subsidence; (15) regions containing extensive Quaternary basalts; (16) Mesozoic and Cenozoic arches and ridge-shaped folds; (17) trend of principal faults, box folds, horsts, and grabens of Mesozoic and Cenozoic age.

ments shows traces of inversion. The Ordos area, for example, was an antecline throughout almost the entire Paleozoic but became a syncline beginning with the Permian. In exactly the same manner, Szechwan was an antecline during the entire Paleozoic and was transformed into a deep syncline in the Mesozoic. Marine deposits accumulated on the platform of China only in the Paleozoic; all the Mesozoic and Cenozoic deposits are continental. Subsidence was always more intensive in the southern part of the platform (south of the Tsinling Range) than in the northern part, where more gaps appear in the stratigraphic section: deposits of the Gotlandian, Devonian, and Lower and Middle Carboniferous are missing.

The development of the platform clearly reveals the course of the Caledonian and Hercynian cycles, both of which begin with subsidence and end with uplift and show the following succession of sequences: Lower and Middle Cambrian, primarily terrigenous, Upper Cambrian and Ordovician, primarily limestone; Silurian, terrigenous (missing in northern China); at the base of the Devonian transgressive series, terrigenous, with limestones above in the rest of the Devonian and the Carboniferous (coal-bearing deposits in northern China); in southern China, in the Lower Permian, limestones; Upper Permian, coal measures; Triassic, lagoonal formations with dolomites and salts; Rhaetic, red molasse deposits; in northern China, the Upper Carboniferous and Lower Permian, lacustrine deposits; Upper Permian and Triassic, coarse molasse.

During the Alpine cycle, the Chinese platform followed a peculiar course of development. In the Jurassic period there was a leveling of the relief by erosion, and extensive lakes were formed. During the Cretaceous, coarse redbeds accumulated; at the same time the platform was broken by faults into a large number of blocks elevated or lowered relative to each other. Many box folds, horsts, and grabens were formed. This tectonic differentiation was accompanied by violent volcanic activity, as well as by the formation (which is unusual on platforms) of great granite intrusives in a zone adjoining the Pacific Ocean in south-east China.

During the Cretaceous period, in that part of the northern China lowland next to the sea and in the region of the Manchurian plain a subsidence of the earth's crust spread from the sea coast into the interior of the continent. In this area of subsidence, however, only continental sediments were deposited (it should be recalled that there was dry land at that time on the sites of the present Japan and North China Seas). During the Paleogene, the crustal movements became less intensive, and lacustrine sediments were deposited in various places. The subsiding littoral plains and accompanying accumulation of continental deposits continued to move inward into the continent. The Neogene and Quaternary periods were characterized mainly by a combination of

two processes: intensive subsidence of the Manchurian plain and the northern China lowland, and uplift of the mountain region in the interior of the continent. This latter uplift encompassed the Tsinling, Nan Shan, Kunlun, and Sikan Mountains and the entire territory of Tibet and the Himalayas. This was a process of "reactivation," which in geologically recent times has embraced an enormous area in central Asia. Subsidence of the plains in the east was, possibly, part of a more general subsidence of the whole vast Pacific Ocean basin—or, more specifically, the beginning of a geosynclinal subsidence in the zone of the Japan and North China Seas.

The intensive activity of the tectonic movements (tectonic differentiation) on the Chinese platform in the Mesozoic, the development of intermediate box folds, the presence of granitic intrusives, and a certain degree of inversion and displacement of the anteklises and synclises all suggest that the Chinese platform underwent a peculiar development approaching what has generally been called parageosynclinal.

It may be that this peculiarity is, to some degree, either directly or indirectly related to some effect of the formation of the Pacific Ocean basin, which subsided considerably sometime after the beginning of the Cretaceous period. Perhaps this subsidence, which was enormous both in area and in depth, caused some extension of the earth's crust in the adjacent regions and thus produced a vertical displacement of blocks of the crust along faults, with manifestations of volcanic activity and the formation of granitic intrusives.

The great depression of the Takla Makan desert and the Tsaidam basin were evidently formed on the site of the intermontane basins existing at the end of the Hercynian cycle, which, in turn, had replaced either the large Hercynian intrageanticlines or small platform areas such as, for example, the Hungarian depression. The Paleozoic deposits in the Takla Makan desert are thin and the top of the Precambrian basement is close to the surface (as in the case of the Tarim massif).

Eastern Transbaykal and the adjacent parts of Mongolia are an Alpine geosyncline closed on all sides, whose principal folding is Mesozoic. This is a small remnant of the extensive Hercynian geosyncline that occupied a vast territory in central and eastern Asia. During the Alpine cycle there was apparently no direct connection between the Transbaykal and the Verkhoyansk or Primorje geosynclines. In the Amur basin these were separated by an area of different and less intensive (or parageosynclinal?) conditions.

Within the above-mentioned geosynclinal and parageosynclinal are that encircles Asia in the south, east, and northeast is an Alpine platform; this is divided into two portions: a reactivated and a nonreactivated part. The first includes all the Tien Shan, the Dzungarian Ala Tau, the Altay, the western Sayan, the Gornaya Shoriya, the Salair, the

Kuznetskiy Alatau, the eastern Sayan, and the Khamar-Daban regions. This is distinguished by its strong and contrasting vertical movements and its corresponding relief. Nevertheless, its previous history was that of a platform divided into relatively gentle subgeosynclines and subgeanticlines.

Both the activated and nonactivated portions of the Alpine platform in Asia rest on a folded basement complex of various ages. The part of the platform between the Urals and the Yenisey River, including the Taymyr, the western Siberian lowland, central Kazakhstan, the Altay, the Salair, the western Sayan, and northern and central Tien Shan, overlies Caledonian and Hercynian folded basements. The Caledonian and Hercynian histories of this area have been discussed earlier. East of here a large part of the platform lies on a Precambrian folded basement. In his structural diagram A. D. Arkhangel'skiy showed a Caledonian basement zone extending from the Sayan, across the Baykal region, to the mouth of the Vilyuy River. The folds in the Baykal region, however, are not continuous, but rather intermediate and discontinuous. There is no igneous activity. From its development this area might be classed as a parageosyncline; but since the scheme presented here does not recognize a parageosyncline in the Caledonian cycle, this area is included in the platform.

The nonreactivated portion of the platform is divided into a number of large anteklises and syneklises. Among the primary anteklises of the first order are the Ural, central Kazakhstan, Taymyr, Anabar, Baykal, and Aldan anteklises. The largest syneklises are the western Siberian, the Tunguska, and the Vilyuy. The western Siberian synecise grades into the Turgay synecise in the south, and thence into the Transcaspiian Alpine subgeosynclinal basin.

These anteklises and syneklises are structural forms of the first order; they are all, without exception, complicated by uplift and subsidence of lesser orders, as illustrated by the structure of central Kazakhstan and the western Siberian lowland. Discontinuous folds are developed over the western Siberian, Tunguska, and Vilyuy syneklises, especially over the last. Over the Transcaspiian synecise, the great domical uplifts of the Mangyshlak, the Tuarkyr, and the Greater and Lesser Balkhan were formed during the Alpine cycle. The Baykal antecise was the site of intensive fracturing and the formation of many grabens. Strong jointing, blocks formed by normal faulting, and extensive younger volcanic activity are characteristic of the Aldan antecise.

The reactivated portion of the platform is subdivided into anticlinoria and synclinoria, as exemplified in the ranges and basins of the Tien Shan and in southwestern Siberia. The line drawn between the reactivated and nonreactivated parts of the platform is tentative, however, since there is a zone of transition between them (see the map at the end of this book).

AFRICA

Nearly all the African continent lies within an Alpine platform (Anderson, 1936; Fourmarier, 1928; Hume, 1935; Krenkel, 1925-1934; Little, 1925; Reed, 1921; Solignac, 1929-1930). The exception is the extreme northwest, where the Atlas Mountains belong to the Alpine geosyncline, which is bordered on the south by a foredeep.

Nearly all the African platform, in turn, lies on a Precambrian folded basement; only in a few places is this basement of a different age. A narrow zone of Hercynian folding parallels the southern foothills of the Atlas. Still farther south, as far as the central Sahara (the Tuareg massif), lies a zone of weak Caledonian folding. The Cape Mountains, at the southern tip of the continent, are an area of Paleozoic rocks folded after the end of the Permian but sometime before the beginning of the Late Triassic. This is, undoubtedly, the northern margin of a vast folded zone whose main body is submerged under the Atlantic. A foredeep at the foot of the Cape Mountains is filled with Late Permian and Early Mesozoic sediments. It has been shown that the periphery of a folded area is the site of the terminal phases of folding; it follows, then, that this post-Permian phase is the end of its cycle, and it is thus clear that the cycle itself must be Hercynian. Hence the Alpine platform at the southern tip of Africa overlies a Hercynian basement. The body of the platform, including Arabia but not the Oman Peninsula with its Alpine folding, lies on a Precambrian basement complex.

The African platform contains a number of synclises and anteklises. The largest are the Karroo (in the foredeep of the Hercynian geosyncline), the Kalahari, the Congo, the Chad, and the Libyan-Egyptian synclises. They are all filled with thick Upper Paleozoic, Mesozoic, and some Tertiary deposits, chiefly continental. Among the anteklises, the most important are two large ones trending north-south along the eastern coast of the continent: the East African and the Eritrean anteklises. These contain the famous grabens, which have been previously described. It should be noted that discontinuous box folds are encountered in the Libyan-Egyptian syncline.

NORTH AMERICA¹

The Alpine geosyncline and folded zone extend along the Pacific coast of North America (the cordillera). The entire central and eastern part of the continent is an Alpine platform.

Crossing the cordillera from east to west, along the latitude of 40° to 45°, one first encounters the Rocky Mountains (see map), a large part of which is a transitional zone from the platform to the geosyncline. To

¹ See Clark, 1930; King, 1933; Reed, 1926; Ruedemann and Balk, 1939; Schuchert, 1930; Williams, 1914; Eardley, 1951.

the east is a subgeosyncline with isolated discontinuous folds in the form of either small domes or isolated sharp uplifts (the Black Hills). The main body of the Rocky Mountains lies in a parageosyncline: here one finds massive boxlike uplifts separated by flat synclinal basins. The igneous activity is represented by small accompanying fissure intrusives. Along the western rim of the Rockies (the Wasatch Range), the linear folds, which are here of the continuous type, form a fan-shaped pattern, with overturning and thrusting both eastward and westward. Only in this zone did true geosynclinal conditions prevail. This was the site of the Rocky Mountain geosyncline, the axial portion of a great area of subsidence whose other parts were relatively quiescent. The axis of the subsidence is asymmetrically closer to the western flank.

The folding in the Rockies is Laramide. West of the Wasatch Range, the so-called Great Basin extends along the axis of the cordillera; this region is depressed in relation to the adjoining ranges (the Wasatch in the east and the Sierra Nevada in the west). The Great Basin is actually composed of many closed depressions, one of which contains the Great Salt Lake.

The Mesozoic and Cenozoic sections here are thin and incomplete. This area is a large intrageanticline that subsided somewhat in the Mesozoic, its subsidence being interrupted by frequent uplifts. At the end of the Mesozoic the entire zone was uplifted and intensively fractured into blocks displaced vertically relative to each other along normal faults. These faults became vents for enormous lava flows. Vast effusions, varying in composition, continued throughout the Tertiary and most of the Quaternary. Southward the Great Basin intrageanticline first expands to form the so-called Colorado plateau, which bulges toward the platform and pinches the Rocky Mountain geosyncline; then, at the latitude of the northern end of the Gulf of California, it wedges out. South of this latitude the Rocky Mountain geosyncline widens and the intensity of its movements increases in its continuation, the Mexican geosyncline (Rellum, Imlay, and Kane, 1936). The subsidence of the Alpine cycle took place here during the Jurassic and Early Cretaceous. After the beginning of the Late Cretaceous, an uplift appeared in the central parts of the geosyncline; the expansion of this uplift displaced the foredeep to the east and northeast. The folding within this geosyncline is Laramide.

West of the Great Basin intrageanticline lies the Sierra Nevada geosyncline. Its history is peculiar: the Alpine subsidence took place during the Triassic and Jurassic, with the inversion, folding, and uplift occurring as early as the end of the Jurassic. Enormous batholiths were formed at the same time. Basins appeared at the foot of the Sierra Nevada central uplift. The eastern basin was poorly developed and drew only a small part of the margin of the Great Basin intrageanticline

into its subsidence. Death Valley is part of this marginal basin. To the south, the marginal basin is better developed. Its submarine continuation is the Gulf of California, while Baja California is the southern projection of the Sierra Nevada central uplift.

At the western foot of the Sierra Nevada central uplift a considerably more pronounced basin was formed, which is topographically represented by the Great Valley of California, famous for its oil resources. This is an intermontane basin. Still farther west there was undoubtedly another central uplift, which has been cut off by the ocean. Its only remnant still standing is the Coast Range, composed of Early Mesozoic and Paleozoic rocks. The California intermontane basin subsided during the Cretaceous, Tertiary, and Quaternary; its sediments are folded into numerous discontinuous brachyanticlines and domes. One of its interesting features is the presence of a great number of vertical faults, both strike-slip and normal faults, mentioned previously.

The northern and central parts of the North American platform lie on a Precambrian folded basement. Different conditions are observed in the southeast, the east, and the extreme north.

Along the Atlantic coast is the Appalachian folded belt of Hercynian age. Its history has been variously interpreted by different investigators. The American geologist Schuchert, for example, has used the Appalachians in his general theory of geosynclines. Schuchert maintained that the continents and the oceans are tectonically different and separate parts of the earth's crust, with different structures and courses of development. In his opinion, the geotectonic zones distinguished on the continents do not continue beneath the oceans. Geosynclines follow the margins of the continents. Hard by the coast is an elevated zone, the *borderland*, with a geosynclinal basin parallel to it on the continental side. The development has two stages: first, a subsidence of the geosyncline, with a deposition of material coming from the borderland; during the second stage, the geosyncline is folded and uplifted, and the borderland sinks. In the meantime, a foredeep is formed along the interior side of the uplifted geosyncline. Ultimately the geosyncline differentiates into foredeep, mountain uplift on the site of the original geosyncline, basin in the place of the former borderland, and ocean. The development of a geosyncline, according to this scheme, is an asymmetrical process.

Over the eastern United States, vertical movements at the end of the Hercynian cycle are reflected in the present topography: in the west, the foredeep of Pennsylvania (Fig. 313) filled with Middle and Upper Carboniferous deposits; east of this, the Appalachian Range uplift, composed of Paleozoic rocks, through and including the Lower Carboniferous; still farther east, the Piedmont plateau, sloping gently toward the ocean and slowly subsiding, so that Mesozoic and Tertiary

sediments lie unconformably on the eroded surface of strongly metamorphosed rocks, which are mostly Precambrian, though some are apparently of Early and Middle Paleozoic age. Schuchert suggested that the Piedmont is the sunken land of "Appalachia," bordered on the west by a Hercynian geosyncline that has since been uplifted.

More recently a number of authors have succeeded in demonstrating the error of Schuchert's scheme (van der Gracht, 1938). It appears that the "borderland" (Piedmont zone) of the American geologists is nothing more than the sunken part of a central uplift—an interior depression. It also has been shown that the structure of the Appalachian geosyncline is similar to that of the European geosynclines, so that its development should also be similar. The only peculiarity in the geology of the eastern part of the U.S.A. is that only the western and the interior portions of the Appalachian geosyncline are visible on the continent, since its eastern flank is covered by the ocean.

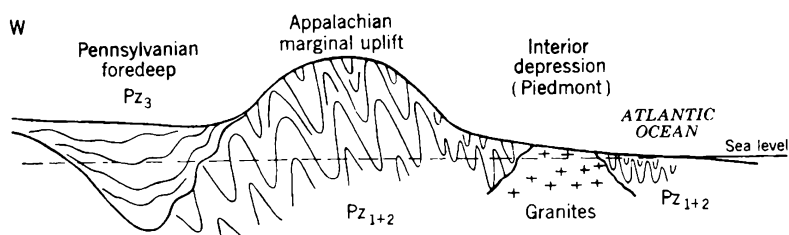


FIG. 313. Diagram of the structure of the Appalachian geosyncline at the end of Hercynian cycle.

Thus the Appalachians are the margin of a central uplift, whose relationship to the Piedmont is the same as that of the Urals to the Turgay "strait." The Hercynian development of the Appalachians is thus as follows: at the beginning of the cycle, subsidence occurred over the Piedmont zone and subsequently spread both east and west. An inversion, apparently taking place—as it did in the Urals—at the end of the Early and beginning of the Middle Carboniferous, led to the formation of a central uplift on the site of the present Piedmont; simultaneously a foredeep was formed over the present Appalachians. In the course of its expansion, the central uplift incorporated the present mountain zone in the west; in the east, it expanded toward the Atlantic. The foredeeps were symmetrically displaced outward. At the very end of the cycle a subsidence of the central uplift took place, with the formation of the Piedmont interior depression. The western part of the central uplift was preserved in the uplifted Appalachians; the remainder of the geosyncline was submerged beneath the waters of the ocean. The preserved part of the Appalachian central uplift subsequently entered into the structure of the Alpine platform as a subgeanticline, and the Piedmont

subgeosynclinal zone lies on the former interior depression. The uplift borders the Gulf of Mexico on the north; the latter is both a part of the terminal Hercynian interior depression and an Alpine subgeosyncline.

Somewhat different conditions prevail in the northern Appalachians, where the geosynclinal development and intensive folding of the Caledonian cycle ended unusually late in the Middle Devonian. During the Hercynian cycle (Late Devonian and Carboniferous), the development in this region was intermediate and parageosynclinal, involving intermediate folding and minor intrusions. The relationship between the Caledonian northern Appalachians and the Hercynian central and southern Appalachians is interesting. The former are not a continuation of the latter, as usually shown on tectonic maps: from geologic maps one can see that the Caledonian folds of the Northern Appalachians crop out on the Atlantic coast in the south and undoubtedly continue under it, "behind" or east of the Hercynian Appalachians. The interrelationship between these two folded zones of different ages is the same as that between the Caledonian northern and the Hercynian southern Tien Shan. The Caledonian structures of the northern Appalachians continue to the eastern coast of Greenland.

The part of the North American Alpine platform that overlies a Precambrian folded basement contains a number of anteklises and synclises. The synclises are the Cisappalachian, corresponding to the fore-deep of the Hercynian Appalachian geosyncline; the Michigan basin; the Illinois basin; and the Midcontinent basin. The anteklises are the Canadian shield, the Nashville dome, the Cincinnati arch, and the Adirondack and Ozark domes. Discontinuous folds, for the most part gentle, are widespread over the synclises. The larger anteklises, notably the Canadian shield, are intensively broken by fractures (bending fractures) and in places by normal faults. The great strike-slip faults within this anteklise have already been mentioned.

CENTRAL AND SOUTH AMERICA²

The Alpine geosyncline, in its passage from Mexico south to the South American Andes, forms an eastward bulge in Central America. The geosyncline here is an oval with the Caribbean in its center. The Greater and Lesser Antilles, as well as the Andes of Venezuela surrounding the sea, form a marginal uplift, and the Caribbean is an interior depression. This is the difference between this sea and the Gulf of Mexico: the latter is characterized by its shallower depth and its smaller positive gravitational anomalies. Tectonically, the Gulf of Mexico is a subgeosyncline.

² See primarily Gerth, 1932-1935, 1955; Rellum, Imlay, and Kane, 1936; Schuchert, 1935.

In South America, the Alpine geosyncline and zone of folding parallel the Pacific coast. The development of the Andes geosyncline has certain peculiar features. It is possible to distinguish three parts of the geosyncline, which together form a single zone yet each of which has a different history: the northern part including the Venezuelan and Colombian Andes, the middle zone of Peru and northern Chile, and the southern part encompassing southern Patagonia and Tierra del Fuego.

In the middle part, in the Alpine cycle, subsidence and deposition began in the Late Triassic and continued to the end of the Jurassic. During that time, the northern and the southern parts of the Andes were regions of uplift and erosion, a continuation of the regime that had existed since the end of the Hercynian cycle. The distribution of the Triassic and Jurassic facies over the central Andes is asymmetrical, indicating a progressive transgression from the west (from the Pacific Ocean) and an exclusively eastern source of terrigenous material (the Brazilian platform). At the beginning of the Cretaceous the subsidence gave way to uplift, which, like the subsidence, expanded from west to east, with a foredeep at its eastern foot during the Neocomian. At the same time there was intensive folding here, with an eastward overturning of the folds and the formation of large batholiths.

Only with the beginning of uplift in the central Andes did subsidence develop over the northern and southern parts of the geosyncline. In Venezuela and Colombia as well as in southern Patagonia and Tierra del Fuego, the subsidence started at the beginning of the Cretaceous and lasted to its end, when an inversion took place in both parts, along with uplift, folding, and intrusions.

The South American Andes are the eastern margin of a complex central uplift. The interior of the geosyncline must now be under the waters of the Pacific. In Tierra del Fuego the uplift turns eastward to form the arc of the South Sandwich Islands connecting South America and Antarctica. This island arc is similar to the Antilles. Within it, judging from all the data, lies an interior depression dating from the end of the Alpine cycle and identical to that of the Caribbean Sea.

There is no doubt that the Andes marginal uplift is very complex, and is divided into a number of uplifts and intermontane basins. This is especially noticeable in the north of the continent (in Venezuela and Colombia) where the Andes, in their descent toward the Caribbean, are separated by broad depressions into a number of ranges. The ranges were elevated at the end of the Eocene, and the depressions have been subsiding ever since the beginning of the Tertiary.

The larger part of South America is located in the area of an Alpine platform. According to the age of its folded basement, it may be divided into two parts, one with a Precambrian and one with a Hercynian basement. Within the Alpine platform, the Hercynian folded zone is exposed

in Argentina. Here, south of Buenos Aires, a wide (up to 700 km), strongly folded Paleozoic zone extends from the southeast to the northwest, plunging beneath the Andes; it is overlain by undisturbed Tertiary deposits. South of this zone (in southern Patagonia), is a large area of Triassic and Jurassic flows resting on strongly metamorphosed rocks. The Hercynian folded zone also includes the Falklands, where Permian sediments lie undisturbed upon eroded and folded Devonian rocks.

The South American Alpine platform is divided, like the others, into anteklises and syneklises. From north to south they are the Guianan anteklise with Precambrian outcrops at its center, bordering in the north on the foredeep of the Orinoco basin; the Amazonian syneklise, with Silurian, Devonian, Carboniferous, Permian, and Tertiary deposits overlying Precambrian rocks (in the west, this syneklise merges with the Cisandes foredeep); the Brazilian anteklise, similar to the Guianan but larger; the Rio Paraña anteklise, where Devonian, Carboniferous, Permian, and Triassic deposits are solidly covered by Rhaetic diabase flows up to 600 thick.

Several smaller syneklises containing Mesozoic and Tertiary sediments, and several anteklises separating them, are known to exist in the part of the Alpine platform resting on a Hercynian folded basement complex.

AUSTRALIA

An Alpine geosyncline and folded zone extend from the islands of the Sunda archipelago across New Guinea, the Solomons and New Hebrides, and New Caledonia to Fiji, Tonga, and the Kermadec islands toward New Zealand.

The continent of Australia, as well as Tasmania, lies within an Alpine platform. A zone of Hercynian folding is located along the eastern margin of the continent. The remaining and greater part of the Australian platform lies on a Precambrian folded basement complex.

CONCLUSION

In concluding this brief review of the tectonic structure of the earth, it should be noted that the general principles of geologic structure, both in the geosynclinal zones and on the platforms, are universal: geosynclines, toward the end of a cycle, are divided into foredeeps, central uplifts, and intermontane basins. Anteklises and syneklises are formed over the platforms -the former with a predominance of fractures and folds and the latter with discontinuous folds.

In some cases, an intermediate parageosynclinal regime develops along the periphery of a geosyncline, implying a less abrupt transition

from the geosyncline to the platform. It has already been observed that many tectonic movements occur in the same succession, in geosynclines as well as platforms.

The general evolution of the earth has up to now been from geosynclines to parageosynclines to platforms. The most important feature is the uneven course of this evolution: its rate has differed in different parts of the earth. This last circumstance is the reason for the simultaneous existence of diverse geotectonic conditions over the surface of the earth. In other words, it is the cause of the whole diversified tectonic structure of this planet.

PART NINE

THE CAUSES OF GEOTECTOGENESIS

CHAPTER 37

THE PRINCIPAL GEOTECTONIC HYPOTHESES

It was said at the very beginning of this book that geotectonics strives for a theory to explain the entire complex of tectonic processes and their succession and interrelationship. In its final form such a theory should be quantitative.

The construction of any such theory, however, let alone a quantitative one, presents problems to which even approximate solutions are not yet in sight. An understanding of the processes within the crust of the earth is impossible without a knowledge of the chemical and physical properties of the material composing the depths of this planet. We do not as yet possess such knowledge. Most important, the mechanical and physicochemical environment of tectonic processes (even within the limits of the earth's crust) is practically unknown. Only the latest high-pressure laboratory experiments have shed some light on that problem. Finally, not all the geologic data available are clear as to their bearing on the sequence and interrelationship of events.

Thus far one can speak only of *hypotheses*, or suppositions, attempting to explain a given tectonic phenomenon and verified by observation. A hypothesis is in no way the crowning glory of a science, as has often—and erroneously—been thought. A hypothesis is essentially a methodologic tool of investigation without which the researcher would be lost in a maze of the phenomena surrounding him. “A *hypothesis*, to the extent that it is correct, is a form of development of natural science” (F. Engels).¹ Of course, hypotheses are of different importance and scope. Many minor specific hypotheses are used in routine scientific work on current problems. In approaching a synthesis of geotectonics, however, a generalizing hypothesis is necessary, and it must suggest an explanation for the causes and interrelationships of all the observed processes. Only such a hypothesis will extend and deepen the synthesis of geotectonics, making it possible to approach the true causes and the mechanism of tectogenesis.

In the first place, any supposition regarding the succession and interrelationship of events must be solidly based on facts and observations.

¹ F. Engels, *Dialectics of Nature*, Gospolitizdat, 1952, p. 191.

In the second place, the hypothesis must embrace as many as possible of the known facts and relationships. The more facts there are that do not "fit" a hypothesis, the less useful it will be. A "neutral" hypothesis—that is, one unaffected by either the presence or absence of any set of facts—is no more acceptable. If possible, all the known phenomena should find their proper place in a stated proposition. In the third place, a hypothetical explanation of tectonic processes should be on the level of present-day scientific knowledge. Any explanation ignoring the phenomenon of radioactivity, and thus standing at the level of nineteenth century science, could not be deemed acceptable in the twentieth century. Finally, a hypothesis must point out the direction of further investigation and therein find its own confirmation. This is a decisive requirement, which is all too often neglected. This means the ability of the hypothesis to predict events: to foretell such new events, relationships, and interdependences as have not been observed before, either in general or in any given case, and which are not directly connected with the basic premise of the hypothesis.

As an example, one may cite a particular hypothesis associating the absence of folding in any area with the rigidity of the earth's crust there. Of course, we have every right to connect the absence of plastic deformations with rigidity of the crust; but how does this promote the further development of the investigation? Every place where there is no folding is called rigid. Rigidity, then, is defined as the absence of folds, and the absence of folding is defined as rigidity. Thus the phenomenon and its explanation are involved in a vicious circle. Such a hypothesis would be justified if it were possible to step out of the circle and verify it by some other phenomenon independent of folding. This would be possible if the rigidity of the earth's crust were manifested in some way other than the absence of folding. But this other manifestation, although there undoubtedly is one related to the property of rigidity, has not yet been found. Thus our hypothesis has led us to substitute "rigidity" for "absence of folding" without suggesting any new predictions or directions for further study.

Altogether different is the supposition that normal faults are a result of tension in the earth's crust. Such a hypothesis is verifiable: if it is correct, then such faults should be confined to areas of uplift, which is indeed the case. Here there is an element of prediction subject to subsequent observation.

The existing geotectonic hypotheses must be evaluated on the basis of the requirements listed above. In so doing, however, it should be kept in mind that any hypothesis is originated in a particular period and at a given level of knowledge. Now, in the middle of the twentieth century, we are right in rejecting the contraction hypothesis in its application to folding; but it would be extremely superficial to consider

it as the product of idle fancy merely because it does not satisfy modern data. There was a time when the contraction theory appeared to be a very progressive conception, fully in accord with the facts then known. Although it is obsolete, its significance in the history of geotectonics should not be underestimated.

Moreover, the complexity and variety of geotectonic processes should be kept in mind, and it should be remembered that for various reasons the attention of different investigators is directed along different paths. We have already seen how geologists were first attracted to the idea of vertical movement of the earth's crust and then to that of folding, only to redirect their attention to the former. Such changes in thinking, of course, were immediately reflected in the nature of geotectonic hypotheses. This is constantly being repeated on a smaller scale: a hypothesis is propounded by an author to explain a certain set of facts that attract his attention and that he believes to be of fundamental importance; another author puts forth another hypothesis because his attention is attracted by another set of facts. The diversity of geotectonic hypotheses reflects both the general evolution and expansion of geologic knowledge and the complexity and contradictory aspects of the tectonic processes themselves. Herein lie the weakness and the strength of the several hypotheses: they focus one's attention on a certain aspect of the tectonic processes, but they do not comprehend the whole, although they evaluate these processes and promote a more complete understanding of them. It follows that any evaluation of geotectonic hypotheses is a complicated problem and should be approached with care.

Hypotheses that are obviously obsolete and only of historical interest are not discussed here. We shall begin with the hypothesis that for decades reigned over geologic science and still has many adherents: the contraction theory. In considering the several hypotheses, we shall not follow a chronologic order. The order of exposition will be dictated by the logical rather than the chronologic relationship between the individual hypotheses. Their respective contents definitely reflect the general development of the science. But their evolution is uneven: some hypotheses appear in advance of their time, often gaining recognition only after several decades, whereas others appear to lag behind. The several hypotheses will be presented in the order in which they would have appeared if they had been "timely," as if each had appeared when the conditions were exactly "ripe."

THE CONTRACTION HYPOTHESIS

The contraction hypothesis is based on the assumption that at one time the earth was a molten globe. It is supposed that, in the course of its subsequent cooling, the earth underwent contraction, its surface

cooling at a faster rate. Near the beginning of the geologic history of the earth, at the time when the solid crust was formed, the outer portion had cooled to the level of the surrounding medium and its contraction ceased, while the subcrustal body of the planet retained its heat and continued to contract during the entire course of geologic history; hence the upward heat flow that is manifested in the form of the geothermal gradient.

Thus the subcrustal portion of the earth continues its contraction. The earth's crust becomes too large for the shrinking body of the earth and hangs loosely over it. Under these conditions, tangential stresses are developed in the crust. According to the contraction hypothesis, these stresses are the principal tectonic forces, which bring about both folding and oscillatory movements as well as faulting and the rise of magma to the surface. All these movements are caused by the decrease in the area of the earth's crust. This shrinkage of the surface causes its folding, and the periodicity of folding is explained by the capacity of the crust to sustain the accumulating tangential stresses for some time, until a sudden collapse occurs. The same shrinking of the surface leads to warping in the form of oscillatory movements, and magma is squeezed outward.

Tangential compression of the earth's crust is the basic premise of the contraction theory. Tangential forces are assumed to be universal; they uniformly embrace all the earth's crust and are of the same magnitude everywhere. But inasmuch as tectonic processes take different courses in different parts of the earth—for instance, each periodic stage of folding affects only a part of the surface rather than all of it—the contraction hypothesis is amended by a conception of the mechanical heterogeneity of the earth's crust manifested in more rigid and more plastic parts. The forces are the same, but the material is different. The rigid areas, instead of being compressed, merely transmit the stresses to the plastic zones, where the compression actually occurs. Each such plastic zone is thus crushed as if in a vise between the rigid "blocks."

The adherents of the contraction theory have produced a conception of stabilizing and mobilizing (that is, making the earth's crust more mobile) factors (Stille, 1918). The latter are purely mechanical. The stabilizing factors, which make the earth's crust more rigid, are folding, igneous intrusion, and uplift—all carrying the crust into the zone of lower temperatures. The mobilizing factor is the subsidence that brings the crust into the zone of higher temperatures, thus increasing its plasticity. The "rigid massifs," then, are the parts of the crust that, having been subjected to repeated folding and intrusion, remain uplifted. The subsiding areas are the more mobile parts of the crust. The plastic segments are Alpinotypic in their dislocations, characterized by continuous folds, according to the terminology used here, whereas the

rigid segments are Germanotypic, with faults playing a substantial part in their dislocation.

The objections to the contraction hypothesis are varied. Objections based on geophysics include such diverse factors as the existence of radioactive heat and the impossibility of further compression of the interior of the earth, which has already been compressed to the limit of its density. It has been suggested that the magnitude of the earth's shrinkage during geologic time, as calculated from folding, would contradict astronomic data. For example, Heim calculated that the Alpine compression, as estimated from the folding and the horizontal displacement of the nappes, is locally as much as 1:8. Even if we postulate that the folding, which spread in the course of past cycles over the surface of the earth, has reduced this surface by only one-half, the volume of the earth would be reduced to one-quarter.

It is to be noted that such objections are extremely vague and ineffective. If the adverse data were all of this sort, the contraction hypothesis would remain unassailable. Indeed, all these objections are based on matters of which there is essentially very little knowledge. To be sure, there is very probably enough radioactive heat to compensate for the constant flow of heat from the earth into space; but this is no more than *likely*, since there are no exact data on the distribution of radioactive elements within the earth. The same is true of the assumptions bearing on the possibility of further compression of the planet's interior. The state of the matter at those depths is unknown; therefore the question of its further compressibility remains unanswered. All that is known is that astronomers recognize densities many thousands of times greater than those encountered on the earth. As regards the amount of the earth's shrinkage as computed from folds and thrusts, all such estimates are extremely inexact. Thus the above-mentioned objections are really not objections at all, but merely doubts that do not contribute to a solution of the problem.

Before the contraction hypothesis can be properly evaluated, our terms must be defined: are we dealing only with the basic premise of the contractionists—their assumption of a gradually shrinking earth—or do we have in mind a contraction mechanism to explain folding and other tectonic movements? This author admits, in principle, the possibility of a gradual shrinking of the earth, owing to its cooling as a result of the migration of radioactive elements within the earth toward its periphery, on the one hand, and to a total decrease in the amount of these elements, because of their decay, on the other. In any case, if such a shrinking has not yet begun, it must inevitably begin in the future. Nevertheless, the earth's contraction has no connection with the origin of either the folding or the oscillatory movements of the earth's crust. True objections can be raised against this familiar application

of the contraction idea in geology, as set forth in the postulates of the classic contraction hypothesis.

It is clear from the foregoing that the oscillatory movements cannot be explained by general tangential forces. A continuing compression does not explain the reversal of the oscillatory movements. If any given area is uplifted as a result of compression, any further compression would only intensify the upwarping; it cannot become a zone of subsidence. General tangential forces, uniformly and constantly pervading the crust, cannot explain the precise action of the oscillatory movements, with their definite sequence and their complexity; nor can they explain the differentiation of the earth's crust into the many regions of very distinct regimes of movement.

It is significant that the contraction hypothesis does not explain the very phenomenon it is supposed to explain: folding. We have seen that the discontinuous type of folding cannot be explained by general tangential forces. But continuous folding does not fare any better; it has been shown above why continuous folds must be considered to be a result of vertical movements of the earth's crust. These objections are likely to be more significant.

A serious discrepancy exists in the explanation of magmatic phenomena. In actual fact, the large intrusions coincide with the time of the folding. But according to the contraction hypothesis, this occurs after the subcrustal part of the earth has contracted somewhat, leaving the crust suspended over it like an arch. Such a situation is most unfavorable for an intrusion of magma into the crust, for the subcrustal matter, because of its contraction, tends to recede from the crust rather than to invade it under great pressure.

Finally, the contraction hypothesis hardly explains the familiar periodicity of the epochs of folding.

Such purely geologic considerations make one categorically reject the contraction hypothesis. The tectonic mechanism implied in it is too rough to explain the refinements and the ramifications of the process of folding and oscillatory movements.

In conclusion, it should be remembered that the contraction hypothesis, in its different stages, has exercised a varying influence on geologic science. The time of its predominance—the second half of the nineteenth century—was an extremely productive period in the history of geotectonics, one of great achievements in both theoretical and practical geology. Great advances were made in the study of folded structures. Therein lies a definite merit of the contraction hypothesis: that it has provided a broad theoretical basis for conceptions of the mechanics of folding. At the same time, however, a theory of geosynclines and epochs of folding was being elaborated, and the first syntheses of the tectonic structure of the earth were presented. Such great authorities in geology as I. Mushketov, Karpinskiy, Borisyak, Obruchev,

Beaumont, Heim, Suess, Dana, and Bertrand have been adherents of the contraction hypothesis.

It should be noted that the interrelationships of various phenomena were recorded correctly in most cases by the adherents of the contraction theory; these have their proper place in our conceptions, although our interpretation is different. An example is the concept of "rigid massifs" as opposed to plastic zones. This conception, so essential to the contraction hypothesis, becomes superfluous if the primary tectonic forces are taken to be vertical, in which case the different intensities of the movement in various regions of the crust are simply explained not by local mechanical differences in the earth's crust but rather by a difference in the intensity of the vertical forces. In the present writer's views, the mechanical heterogeneity of the crust is considerably less important, since it affects only the details of the structure being formed.

On the other hand, the conception of a rigid massif takes on another aspect if it is recalled that, according to the contraction hypothesis, uplift is one of the "stabilizing" factors, whereas the mobility of a plastic zone is determined by its subsidence. Then, translated into more familiar terms, the relationship between "plastic" and "rigid" segments becomes one between the zones of subsidence and uplift, between intrageanticlines and intrageosynclines, and between geosynclines and platforms. This relationship is expressed in the fact that intensive folding is confined to zones of subsidence, in the overturning of the folds away from such zones and toward the uplifts, etc. Hence the relationships remain as recorded in the contraction hypothesis but are given a different interpretation. They are not associated with the different mechanical properties of matter but with differences of a more profound thermodynamic character, manifested in the different magnitudes of the tectonic forces and the different processes of development.

Later on, at the very end of the nineteenth century and especially in the twentieth century, the contraction hypothesis, as we have seen, was of negative value to science, after becoming only a basis for a tendency toward formalism in geotectonics. This was discussed in Chap. 2.

THE HYPOTHESIS OF ISOSTASY

As previously indicated, one of the greatest obstacles in the path of the contraction hypothesis was the need to explain oscillatory movements. The interpretation of these movements was the basic task of some of the other geotectonic hypotheses. Above all, one must mention the attempts to utilize certain isostatic concepts for tectonic purposes. Following the original general works of Pratt (1855) and Airy (1855), which set forth the general theory of isostasy, Dutton's work (1892) played an important part in spreading the idea.

We have seen that the observed state of equilibrium of the earth's

crust is satisfactorily explained by the supposition that the crust is floating upon a denser and more viscous substance. If this is true, the accumulation of sediments over one area and their erosion over another must lead to a subsidence of the former under the weight of the load and to an uplift of the latter because of the decrease in the weight of the crust. It is by this device that some authors attempt to explain oscillatory movements while still adhering to the contraction hypothesis for an explanation of folding. Thus the isostasy theory postulates two directions of the forces instead of just one and two causes for the tectonic movements of two basic types.

It was shown earlier, however, that the accumulation of sediments cannot cause subsidence of the earth's crust (Chap. 13). The relationship between the two phenomena is just the reverse: the subsidence is the cause of the deposition. Moreover, the cause of the oscillatory movements still remains unknown. If an unloaded area rises while a loaded one subsides, there is nothing to bring about the observed reversal of the process and the uplift of the previously subsiding area.

In regard to folding, the isostasy hypothesis repeats and augments the mistakes of the contraction hypothesis: if the earth's crust had been so pliable as to warp easily under a load of sediments, it could never have transmitted tangential stresses over great distances; instead, it would have been folded everywhere, and platforms could not have existed. Furthermore, this hypothesis separates folding and oscillatory movements, attributing them to different causes and thereby destroying the unity of the tectonic process, which, however, is evident from all observations.

Thus at the present time the isostasy hypothesis is even less profitable than the contraction theory. Nevertheless, in rejecting it one must acknowledge its positive influence on geotectonics, inasmuch as it has attracted geologists' attention to the vertical movements of the earth's crust. Isostatic forces and isostatic equilibrium are actual facts. We are not disputing their existence but only stating that their role as geotectonic factors is very limited.

BUCHER'S PULSATION HYPOTHESIS

Another attempt to improve upon the conceptions of the mechanism of oscillatory movements and explain their connection with tectonic movements of other types was that of the American geologist Bucher (1933, 1939). In Bucher's view, the life of the earth is a series of recurrent phases of general dilation and contraction of the earth's crust associated with pulsation of the subcrustal body of the planet. By analogy with the corresponding phases of the heart's action, the phase of dilation and expansion of the earth's crust is called the *diastolic*

phase and that of contraction is called the *systolic phase*. During the expansion of the earth's crust, its more elastic parts undergo a greater extension and consequent thinning. The surface expression of this process is subsidence, with the formation of geosynclinal basins subsequently filled with sediments. In the systolic phase, these plastic areas are folded.

The conditions of the systolic phase are identical to those postulated by the contraction hypothesis; therefore, all the objections to the latter remain in force here. Only the oscillatory movements fit better into Bucher's pulsation hypothesis. At the same time, it introduces new difficulties. To begin with, the very idea of a historical occurrence of general phases of dilation of the earth, tangentially embracing its entire crust, is unrealistic: even if we postulate the crudest scheme of a single subsidence in each geosyncline, the observed lack of synchronization in the corresponding vertical movements in different geosynclines remains unexplained. If we consider the universal and continuing existence of zones of subsidence together with uplift, both within geosynclines and in the zones of subsidence and uplift outside them, the difficulties become insuperable. Indeed, if subsidence is brought about by expansion and uplift by contraction, how can central uplifts, foredeeps, and intermontane basins stand side by side? A peripheral geosynclinal foredeep seems to suggest extension of the crust. But within the extended zone lie areas of uplift and folding areas where, according to this hypothesis, contraction takes place. These elements are irreconcilable.

Like the contraction hypothesis, this theory misinterprets the interrelationship between igneous activity and other tectonic phenomena. Here the rise of magma also coincides in time with contraction, with the maximum "separation" of the subcrustal material from the crust and the minimum magmatic pressure.

The value of this hypothesis lies in the fact that it, more than many others, has directed attention to the periodic character of tectogenesis and associated that feature with the periodic changes in deep-seated subcrustal processes. In principle, this hypothesis is on the right track, but its concrete formulation must be considered to be in error.

THE PULSATION HYPOTHESES OF V. A. OBRUCHEV AND M. A. USOV

The Soviet scientists V. A. Obruchev and M. A. Usov have attempted to improve Bucher's conceptions and to produce a more satisfactory version of the pulsation hypothesis.

M. A. Usov pointed out (1936-1940) that the attempts at geotectonic generalizations outside the Soviet Union verged on metaphysics. While admitting a certain alternation of phases of contraction and dilation,

these foreign authors consider any given process to be an alternation of states, devoid of any internal contradictions. But the life of a cosmic body, according to dialectic laws, is determined by the enduring coexistence of two factors—attraction and repulsion—which are in a constant struggle, with sometimes one and sometimes the other prevailing. The struggle between contraction and dilation occurs throughout the body of the earth, with—as we understand Usov—contraction predominating in the crust and dilation in the depths of the earth. In this struggle between opposites, the leading factor is compression. During the quieter periods of the earth's life, this struggle is manifested in slow, wavelike oscillatory movements.

Usov wrote (1940):

At some stage the change in the physicochemical conditions, at least in some of the levels within the earth, reaches a critical point beyond which the greater part of the material in these levels passes into a different atomic state. This change is undoubtedly rapid and accompanied by a sharp change in volume. If this takes the form of a decrease in volume, related to a corresponding increase in thickness of a denser geosphere at the expense of a less dense mantle, the peripheral part of the earth would contract within a brief length of time, thus leading to tangential diastrophism.

With the latter, Usov associated folding, thrusts, and acidic intrusions. Should there be an abrupt expansion, the earth's crust is broken by fissures through which magma is extruded.

V. A. Obruchev (1940), after a detailed criticism of the views of Bucher, Usov, and other authors, formulated his version of the pulsation hypothesis, which he believed to be in better accord with the principles of dialectic materialism. The content of Obruchev's theory is practically identical with that of Usov's views, except that Obruchev provided a more complete treatment of the formation of magma and of metamorphism in the general scheme of the earth's development.

Obruchev wrote:

It can be accepted that the history of the earth consists of protracted periods of evolution during which the struggle between contraction and expansion does not cease, but merely abates to slow oscillatory movements and to isostatic restoration of the equilibrium disturbed during the times of revolution. The latter, alternating with the periods of evolution, are distinguished from them by their shorter duration and by the greater intensity of the contest between compression and extension, which is resolved abruptly, thus leading to sharp changes in the relief of the earth's surface. Each of these periods, which may be called cycles, consists of a number of phases in which there is a dominance of either contraction or expansion.

In this writer's opinion, there are important errors in the hypotheses of both Usov and Obruchev. While assuming that both folding and intrusions take place upon contraction of the surface of the earth—a vic-

tory of compression over extension -these authors confront the same difficulties that beset all theories that derive folding from external tangential pressure. In its existing form, therefore, and as far as the succession of movements of the earth's crust is concerned, this hypothesis is little different from Bucher's and shares many of its shortcomings. Among these, beside the notion of a tangential pressure causing folding, is the conception of compressed areas within expanding regions, of magma rising during the period of the dominance of contraction, etc. Another important defect is the lack of any physical basis for the processes that Usov and Obruchev assumed to occur in the earth's depths.

THE HYPOTHESIS OF RADIOACTIVE CYCLES

This conception of Joly's (1925, 1929), like Bucher's hypothesis, is based on the periodicity of tectonic processes. It differs in its attempt to provide an explanation for the earth's pulsations, which is lacking in Bucher's work.

Joly begins with the isostatic conception of granitic continental masses floating upon a basaltic substratum. Both the granites and the basalts generate radioactive heat. Some of this heat is dissipated in space, but at a certain depth it accumulates and the temperature rises. All this takes place within the basaltic layer. Most of the heat is accumulated under the continents, which, because of their low conductivity, hinder the escape of the heat. The accumulation of heat beneath the oceans is slower. The heat accumulated over 35 to 50 million years melts the basaltic shell; upon melting, the basalt expands and its density decreases. Hence the continents sink deeper and are partly covered by transgressions of the sea. This expansion of the basaltic layer leads to the formation of fissures in the continents, through which effusions take place.

After the basalt has become liquefied everywhere at a certain depth, the solid crust above it rotates, under the attraction of the sun and the moon, from east to west. The ocean bottom is thereby displaced with relation to the molten basalt into areas previously occupied by the continents. This allows the areas of heated basalt to cool off by losing some of their heat to the outside, through the ocean waters. As a result, the basalt hardens again. Upon hardening and regaining its density, the basaltic layer contracts, the granitic continents are reelevated, regression of the seas begins, and folding follows the compression. Thereupon the entire cycle is repeated. This hypothesis was somewhat modified by Holmes (1927, 1928-1931), but its essential content remained the same.

The physical aspect of Joly's hypothesis became the subject of vigorous criticism by several authors. It was shown that periodic warming and cooling of the substratum are physically inconceivable. A planet

built on Joly's suppositions would unavoidably and rapidly fall into a state of thermal equilibrium. Moreover, Joly assumed that the melting temperature of basalt is lower than that of granite, although in reality it is higher. The notion of the effect of the sun and moon and of the rotation of the outer layer of the crust, which is essential to the cooling of the basalt and the preservation of the continents from becoming molten, is also very artificial.

It would seem that the value of this hypothesis lies in its graphic presentation of the periodicity of tectogenesis. Even here, however, there is an important latent methodologic fallacy: the life cycle of the earth, according to this hypothesis, is a closed circle of countless repetitions of the same events. This cyclic idea hinders the view of a more general progress: the true course of events does not run along an unending circle but moves rather as a spiral. Finally, the purely geologic objections to this hypothesis are the same as those directed against Bucher's.

THE CONTINENTAL DRIFT HYPOTHESIS

The failure of these geotectonic hypotheses, especially the contraction theory, brought forth, chiefly in the period of 1910 to 1930, a flood of various hypotheses, the vast majority of which must be regarded as fantastic and having nothing to do with science. This is true of the hypotheses suggesting horizontal drift of the continents, among them the hypothesis of Wegener (1924), which was once famous.

The basis of this hypothesis is the isostatic concept of granitic continents floating upon a basaltic substratum. Wegener postulated that originally all the granitic material was spread in a continuous thin layer over the surface of the earth. The tidal forces pulling all objects on the earth's surface from east to west, as well as the centrifugal forces causing pressure directed from the poles to the equator, drove this material together into a single massive block, called the *continent of Pangea* (Fig. 314). All this took place during the Paleozoic. During the Mesozoic and Cenozoic, the same forces split the continent of Pangea into separate parts.

America was set free from Europe and Africa and, outracing the rest of the continental blocks in their common westward drift, advanced far ahead, producing the Atlantic Ocean in its wake. Resisted in its advance by the basalt substratum, the front of the American continent was warped into folds, as a result of which the Cordilleras and the Andes Mountains were formed.

Africa was partially separated from Asia, and its southern tip was rotated somewhat, clockwise, to make room for an ocean between it and India, which was formerly adjacent. Antarctica and Australia, split

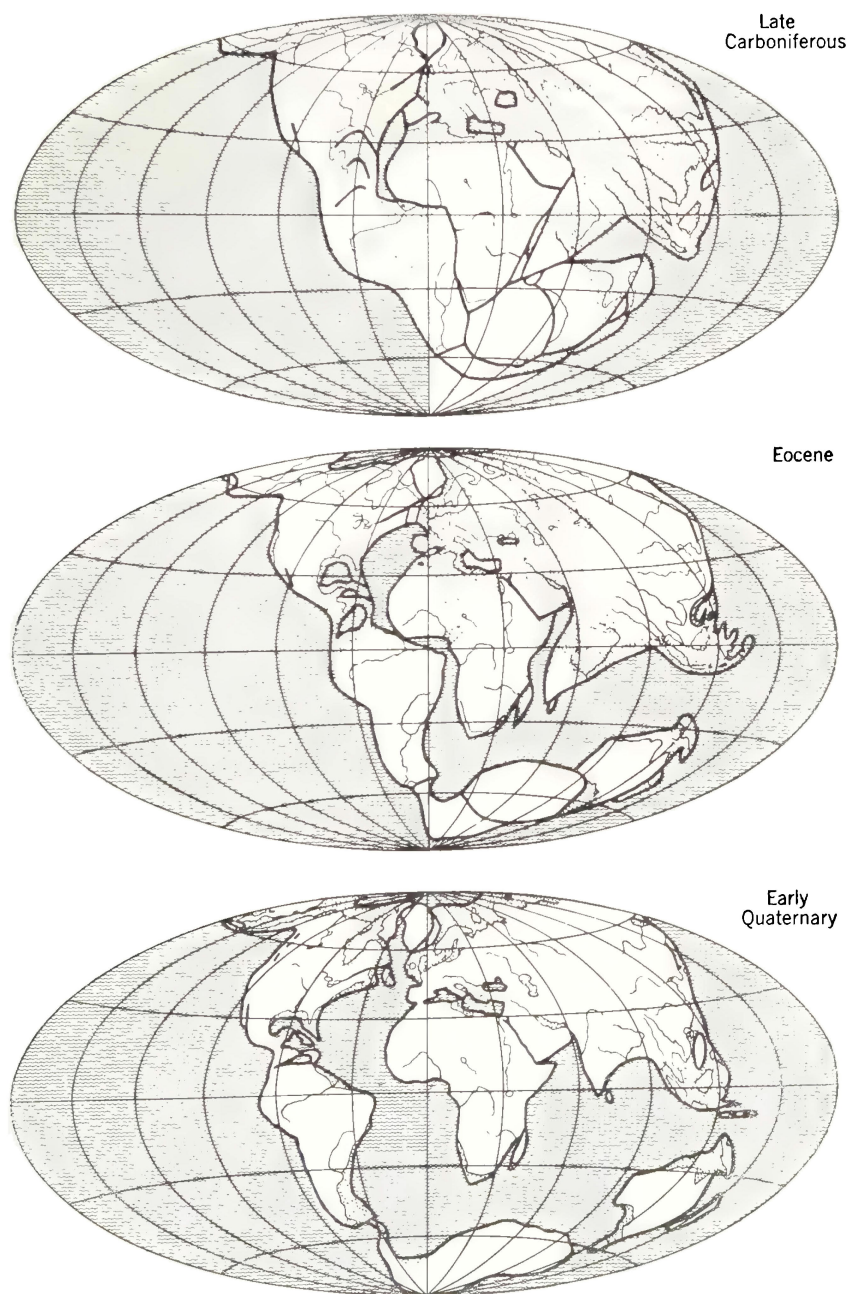


FIG. 314. Continental drift, according to A. Wegener.

apart from Africa and Asia, drifted southward and then away from each other.

The folded zone extending parallel to the equator from the Mediterranean to the Himalayas was produced by pressure along the equator. This pressure, in turn, was induced in the continents of Eurasia and Africa by centrifugal forces.

The island arcs, girdling the continents on the east (such as the Aleutians, the Kurils and Japan, the Philippines, and Great and Little Sunda near Asia or the Antilles and the South Sandwich Islands near America) are bits of the continents lagging behind in the westward drift. Because of friction against the simatic layer, some of the continents were bent, as in the curvature of Tierra del Fuego.

Wegener supported his hypothesis with various arguments. In the first place, he referred to the geophysical data on the heterogeneity of the continents and the oceans. Secondly, he pointed out that the outlines of many continents are roughly parallel to each other; because of this, it is possible to put the continents together into a single block, like a jigsaw puzzle, by pushing them toward each other and rotating this or that continent at some angle so as to leave hardly a chink between them. Thus the east coast of America almost duplicates the curvature of the west coast of Europe and Africa in such a way that, if these continents are put together, the eastward bulge of Brazil will fit into the Gulf of Guinea; Africa's westward bulge, north of this Gulf, corresponds the curvature of the eastern coast of Central America, etc. The correspondence is even more complete if the triangle of Greenland is wedged in between America and Europe, in the north. A similar parallelism of the coasts was established for the other continents. In reading Wegener's book, one obtains the very clear impression that this coastal parallelism was the main basis of his hypothesis.

Third, Wegener referred to the structural similarity of the continents, although they are separated by oceans. He pointed out, for example, that the geologic structures of South America and Africa are so similar that they might be parts of a single structural complex. In the fourth place, Wegener indicated other similarities between the now separated continents. For instance, a considerable similarity in the development of the flora and fauna of South America, Africa, India, and Australia, up to a certain time, was cited as indisputable testimony of the physical connection between these continents during the Paleozoic and the beginning of the Mesozoic. Moreover, certain paleoclimatologic considerations were voiced. In his attempt to reconstruct ancient climatic zones from the character of the sediments, Wegener concluded that such a reconstruction is more easily achieved on the supposition of a distribution of the continents different from the present one.

It is a source of profound amazement that such a hypothesis—based

as it is on an overtly formalistic approach to the major problems and on a total and consistent disregard of the basic geotectonic data and, as already stated, explaining nothing of what must be explained in the first place—was not only seriously discussed in scientific literature but achieved considerable success and attracted some of the leading authorities into the ranks of its adherents. These men were apparently hypnotized by the boldness of Wegener's ideas and by his brilliant style of writing.

It has been shown that the oceans and continents are one, geotectonically speaking, and that structural zones can be traced from the continents into the oceans. This alone undermines the very foundations of Wegener's hypothesis. The symmetry in the distribution of the centers of stabilization would be only accidental if the continents did move. Such fortuitousness is utterly improbable.

Leaving these major problems for more elementary considerations, one perceives even more clearly the total vacuousness and sterility of the hypothesis. Oscillatory movements are the fundamental type of tectonic movement. Where are these movements in Wegener's hypothesis? Where is the differentiation of the earth's surface into areas of different oscillatory regimes? Where is the regular periodicity of the oscillatory movements, their orderly succession, etc.? Turning to the folding movements, we find that this hypothesis has no room for discontinuous folding. Nor, for that matter, do continuous folds fare any better. The idea of the crumpling of the advancing front of America in its drift over the substratum can hardly be counted as an explanation of the folding. If the substratum is softer than the granitic continent, it is the substratum and not the continent that should be crumpled; on the other hand, a softer continent would not budge against a harder substratum. There is even less chance of finding in this hypothesis any explanations of such phenomena as the periodicity of folding movements, the migration of the phases, and the regularities in the relationship of folding to oscillatory movements. Only extrusive igneous activity is touched on. The entire history of structural development, local as well as that of the earth as a whole, is completely disregarded.

The question arises: what of the arguments in support of these ideas? Under more careful investigation, the direct continuation of the structures from one continent to another is not confirmed. There is some similarity in geologic structures, but nothing beyond the usual similarity of the facies and structures between tectonic zones of the same type. Just as plausible a similarity could be found between the Alps and the Himalayas, which were certainly never in contact. A direct connection between some of the continents, providing a means of faunal and floral interchange, certainly did exist but not necessarily in the drift of one continent toward another. The dry land and shallow seas that might

have existed over the sites of the Atlantic and Indian Oceans have already been mentioned. Such "bridges" are more likely than Wegener's ideas.

The paleoclimatologic considerations cannot be decisive; in any event, they do not outweigh the above objections. Any reconstruction of ancient climatic conditions is inexact; even the principles of a methodology of paleoclimatologic reconstruction have not been worked out.

There remains the parallelism in the outlines of certain continents. Indeed, this is a curious feature; a rational explanation has not yet been found. But is ignorance of its true cause sufficient reason for turning to a fantastic explanation?

Despite their differences, all the hypotheses discussed so far have this in common: taking folding to be of paramount significance among all tectonic movements, they all strive for an explanation of folding. In so doing, they take the formation of folds to be determined by external tangential stresses. Hence an interpretation of these tangential stresses is the main content of these hypotheses. An exception is the isostasy hypothesis, which gives priority to the oscillatory movements but divorces them from folding.

It was seen from the foregoing chapters, however, that the creation of a tectonic generalization in the light of modern data is possible only when all tectonic movements are subordinated to vertical forces and when one takes oscillatory movements to be the main form of tectogenesis. The hypotheses that follow that path are the most progressive at the present time. We may now consider these hypotheses.

THE HYPOTHESES OF CONVECTION CURRENTS

This class includes a group of hypotheses according to which slow currents similar to convection currents arise beneath the earth's crust. These ideas were initially elaborated by Ampferer (1906), who postulated contraction and expansion of the subcrustal matter as the causes of oscillatory movement. When expansion takes place in the depths, an ascending current is formed that uplifts that part of the earth's crust, first forcing it to warp upward and then flowing to the side in horizontal and descending currents. Horizontal currents, because of the friction of the plastic material against the earth's crust, leads to displacement of the latter: the crust is carried along with the deep magmatic current and, encountering resistance, is folded.

Ampferer said nothing of the causes of these subcrustal currents. Other authors, during earlier periods in the history of this hypothesis, have mentioned changes in volume resulting from crystallization, radioactive heat, tidal phenomena, etc. (Andrée, 1914; Noelke, 1924; Kossmat, 1921).

The great interest in the idea of subcrustal convection currents arose

in the fourth decade of this century and has not died out since. Different variations of the convection-current hypothesis are at the present time the most widespread in the tectonic literature of the world. Most recently, geophysicists, rather than geologists, have appeared as authors of the various convection hypotheses. And the principles of these hypotheses themselves are almost exclusively founded on geophysical conceptions of the properties of the earth's subcrustal material and of the thermal regime in the depths of the earth. On the other hand, geologic data on the nature and succession of crustal movements are used in extremely simplified and schematic form. In these hypotheses, for example, geosynclines are invariably conceived as single areas of subsidence, which at a certain stage undergo horizontal compression, as a result of which folding originates. The complexity of geosynclines and the many types of folds are disregarded, and the tectonic evolution of platforms is never taken into account.

Among the authors of different versions of the convection hypothesis are Pekeris (1935), Griggs (1939), Vening-Meinesz (1952), Kraus (1951), and Urey (1953). All the versions produced by these writers assume that the convection currents are due to heating of the material within the earth's depths by heat from radioactivity. The heated masses move upward and are replaced by cooler substance; the material that has risen nearer the surface is in turn cooled, and so forth.

Pekeris has devoted most attention to a mathematical theory of convection currents and to the factors determining the magnitude of the convection cell, the rate of the movements, etc. Griggs has pointed out the possibility that convection may explain the periodicity in folding: if it is assumed that the material in the earth has an ultimate plasticity, a certain amount of heat will be required to cause this material to begin to rise. Since the rise of the heated material and the sinking of the cooled substance take place more rapidly than the earth's loss of heat by heat transfer, after the "overturning" of the convection cell, another period of time will be needed before the cold material that has traveled downward is heated to the point of rising. Griggs also examined and reproduced on laboratory models the mechanism by which the earth's crust is drawn downward, with the formation of a narrow depression, by two convection currents that move toward each other and flow downward when they meet. Ultimately, this area of subsidence is crumpled horizontally.

Vening-Meinesz has suggested that the crumpling of the earth's crust and the formation of a geosynclinal subsidence are due to the horizontal compression produced by two convection currents flowing toward each other and carrying the crust along, compressing it above the area where the currents come together and descend (Vening-Meinesz, 1950).

Kraus believed the chief factor in geosynclinal development to be the drawing downward of the earth's crust in certain zones, with the forma-

tion of areas of subsidence and their succeeding compression. He spoke of two levels of convection currents, an upper and a deeper one, differing in the direction and nature of the currents in them. The movements in the deeper level determine the major structural features of the continents and the oceans, whereas the convection in the upper level causes the development of structural forms on the usual geologic scale.

In the writings of Kraus and especially of Urey, in contrast to those of many other authors, convection currents are considered not as purely mechanical phenomena—as a circulation of material with an unvarying composition—but also from the standpoint of possible phase and chemical changes in the material as it moves through zones of different temperatures and pressures.

The hypotheses of this group contain interesting and useful conceptions. But these hypotheses on the whole, inasmuch as they are based only on geophysical considerations, in their geologic aspect so severely schematize tectonic processes that they are very far from reality in their conclusions.

THE GRAVITY FOLDING HYPOTHESIS

In 1930, Haarman proposed a hypothesis based on the conception that the leading role in geotectogenesis is played by vertical crustal movements, which he called oscillatory. Haarman called oscillatory movement *primary tectogenesis*, and he took folding to be *secondary tectogenesis*. The oscillatory movements are induced by a displacement of the liquid portion of the sialic layer contained under the continents. In this process, the earth's crust is locally uplifted to form bulges (*geotumors*) or downwarped with the formation of depressions (*geodepressions*) (Haarman, 1930).

The slope of the flanks of the bulges is sufficient for the deposits in them—their mobility increased by water—to begin sliding by gravity down the slope. In this process, the strata at the top of the slope are fissured and faulted, and those at the bottom are folded.

It has already been stressed that Haarman's ideas of primary and secondary tectogenesis, of the subordination of folding to oscillatory movements, and of the leading role of gravitational forces in folding are a major step forward. This hypothesis was the beginning of the lines of development of general ideas that now are the core of geotectonics.

THE WAVE HYPOTHESIS

The ideas in Haarman's hypothesis dealing with the causes of uplift and subsidence, as well as with the succession of oscillatory movements,

were developed in greater detail by van Bemmelen (1933, 1955). Van Bemmelen indicated a dual process developing in the earth's crust, destroying the equilibrium of the mass in it and later restoring the equilibrium. The first consists in a differentiation of the silicate mantle as a result of its cooling, and the second in a redistribution of the deep-seated masses, leading to a restoration of this balance.

It is supposed that between the upper (sialic) and the deeper (simatic) layers there is an intermediate layer of mixed composition, in

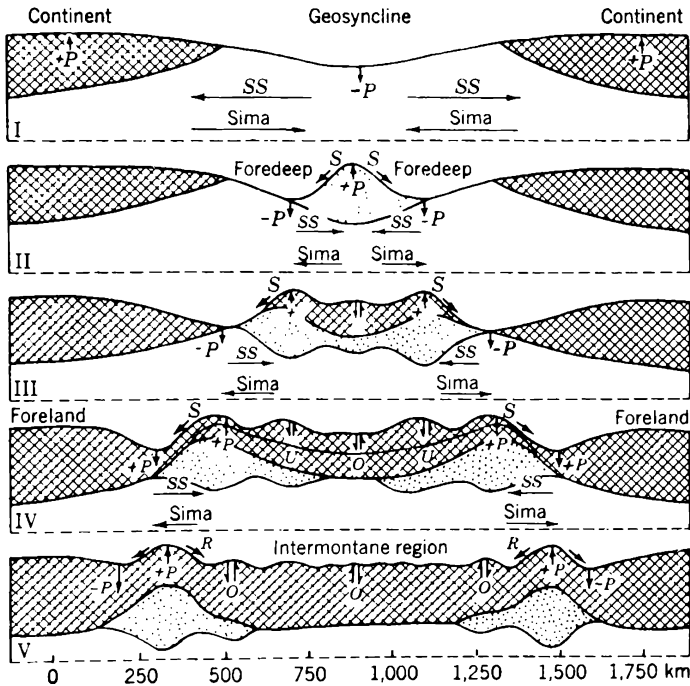


FIG. 315. Scheme of tectogenesis, according to van Bemmelen. (I-V) stages: (I) geosynclinal; (II) embryonic; (III) young; (IV) young maturity; (V) mature; SS, salsima; +P, primary tectonic uplift; -P, primary tectonic subsidence; O, primary tectonic oscillation; S, secondary tectonic sliding of strata; R, secondary reverse sliding; crosshatch, sial; blank, sima; dots, mixed sial and sima.

which a process of magmatic differentiation occurs. As a result, the lighter elements (sial) rise to the top, and the heavier (sima) sink to its base.

In the regions of intense cooling, the magmatic material at depth contracts, and a geosyncline is formed above it. Simultaneously, a differentiation takes place, most intensively under the axis of the geosyncline; new amounts of sial float upward to make a bulge in the earth's crust and to form what we have here called a central uplift (Fig. 315). The latter gradually expands and breaks into two parts, separated by a basin of deposition.

Thus van Bemmelen has proposed that there is a succession of the oscillatory movements within a geosyncline, which is now confirmed by observation. Folding movements he interpreted after Haarman, in terms of a gravitational sliding of the sedimentary strata on the surface down the slopes of the convex bulges in the earth's crust.

A very important aspect of van Bemmelen's conception is its use of the process of physicochemical differentiation as the basic factor in the subcrustal movements. By this means van Bemmelen has differentiated his views from the entirely mechanical conceptions of the authors of the convection hypotheses mentioned above. There can be no question that a satisfactory theory of geotectonics can be developed only on the foundations of a study of physicochemical changes taking place in the material composing the earth's depths.

M. M. TETYAYEV'S VIEWS

A work by M. M. Tetyayev entitled *The Foundations of Geotectonics* appeared in print in 1934 and was reissued in 1941. This book gave rise to much discussion and was greeted variously but for the most part unfavorably. Both the questions that were posed and the argumentation were highly unusual. The present writer believes that here, in the light of all that has been presented in this book, is a suitable place for a more complete and objective evaluation of M. M. Tetyayev's work.

This writer is convinced that the publication of *The Foundations of Geotectonics*, in spite of its many and obvious defects, was an event of the greatest importance for geotectonics. Tetyayev's central idea, the systematic presentation of which was his greatest achievement, is that the different types of tectonic movements are different forms of a single process of geotectogenesis, which, in turn, is part of a still more general process of cosmic development of the planet as a whole. Although the basis of this idea is not new, and may be found, for example, in the hypotheses of the contractionists, Tetyayev was the first to elaborate it so effectively and thoroughly.

Numerous works on tectonics contain much information on the manifestations of different tectonic movements, but no one—at least, until M. M. Tetyayev—had attempted to determine what these manifestations had in common or to discover the regularities in all of them. After the publication of *The Foundations of Geotectonics*, M. A. Usov and V. A. Obruchev, whose ideas have been discussed earlier, set out upon this path.

This posing of the general problem also made possible a new formulation of more particular tasks for research. In addition to studying the specific forms of tectonic movements, attention was concentrated on discovery of the laws uniting these movements and on establishing the connections and relationships between the various phenomena. In-

vestigation of the relationships between oscillatory movements, folding, and other forms of tectogenesis and between such movements and the fissuring of the earth's crust, as well as clarification of the role played by igneous activity in relation to other forms of tectogenesis, have now become the central tasks of geotectonics. It is easy to see that only a solution of these problems can lead to discovery of the general laws of development of the earth's structure. It should not be inferred from these words, the present writer hastens to add, that nothing had been done in this direction until the appearance of Tetyayev's work. Even earlier, investigators had set themselves such tasks as the elaboration of the theory of geosynclines but had carried them out in an elementary manner, according to the limitations of the various data from observations, and attempted to solve the problems only to the extent required by the particular case in which they were interested. These problems had never before been formulated as a fundamental goal of research or in such thorough form.

Posing the problem of the unity in tectonic processes has predetermined much that is of value. Above all, it made possible a considerably more precise formulation of the goal and purposes of geotectonics and of its position among the other branches of geology. In itself, however, this emphasis on the unity of tectogenesis would have been fruitless, if the manner in which this unity is manifested had not also been indicated. The greatest difficulty in uniting the tectonic processes into a single system has been the seeming contradiction between oscillatory and folding movements. Tetyayev suggested the existence of a mechanism that produced folding as a flow of crustal material along the strata, under the influence of forces acting vertically. Although the concrete features of this mechanism were not presented with sufficient clarity, the conception itself was undoubtedly correct.

The great shortcomings of Tetyayev's work were an excessive amount of inference, an insufficient factual basis for its conclusions, and the substitution of such obfuscated phrases as "the struggle between condensation and repulsion" within the earth for concrete physical observations. Tetyayev thought more in terms of abstract philosophical categories than of natural and historical conceptions. But one must not fail to do justice to the force of his generalizing mind, which was able to perceive unity where earlier students had seen primarily disconnected and unrelated phenomena.

THE ASTHENOLITH HYPOTHESIS

In 1941, a paper by B. and R. Willis appeared, entitled "Eruptivity and Mountain Building." The authors began with a statement to the effect that the failure of previous attempts at an explanation of mountain-forming processes was due to ignorance of the source of tectonic

energy. This source, in their opinion, is radioactivity. Radioactive elements are unevenly disseminated throughout the body of the earth: in some places they are accumulated, causing the temperature to rise and melt the material in which they occur. Enormous masses of molten magma (asthenoliths) thus formed begin to rise. It is supposed that they originate deep within the earth (in its core) and then make their way to the surface. As they rise, they pass through constantly changing conditions and therefore undergo differentiation, changing from their spherical shape to columnar. This process gradually substitutes a mosaic of extrusive and intrusive bodies for the primary crust of the earth.

Data show that the schistosity in the Laurentian, Adirondack, and other crystalline massifs was caused by their recrystallization under high pressures and temperatures. Recrystallization is associated with flow of the material and with a decrease in dimensions in one direction and an increase in the two others. In many cases, horizontal schistosity

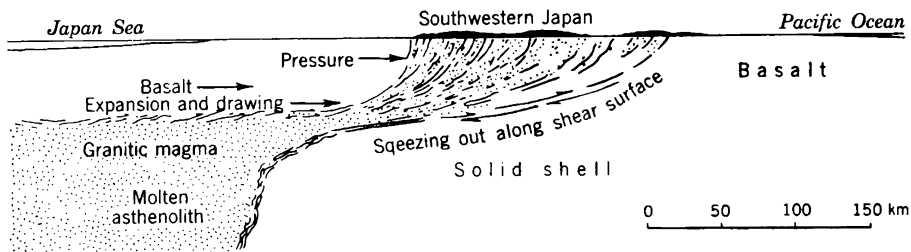


FIG. 316. Diagram of tectogenesis according to the asthenolith hypothesis of B. and R. Willis, as exemplified in the structure of the islands of Japan.

is observed, suggesting horizontal elongation and vertical compression. The elongations attain a magnitude on the order of tenfold.

The physical state in the earth's crust is as follows: the ambient pressure of the weight of the rocks increases the viscosity of the crust; a rise in the temperature decreases it. Generally speaking, the pressure exceeds the corresponding magnitude of the viscosity, probably at the depth of 25 to 35 miles (40 to 55 km); should a pressure differential originate at that depth, the rocks will flow. Moreover, the earth's crust possesses certain elastic properties and thereby resists mechanical stresses.

When a column of molten magma, in its rise from the depths, approaches the crust, its heat melts or else merely softens the rocks at the base of the crust. As a result, these rocks lose their solidity and equilibrium, and the material begins to flow under the weight of the overlying rocks, being squeezed first horizontally and then along a surface slightly curved upward (Fig. 316).

The formation of a column of magma is accompanied by expansion of its material, first as a result of the rise in temperature while it is still

in the solid phase and then because of its melting. The increase in volume should be approximately 5 per cent.

After the melting is complete, the forces become hydrostatic, the pressure is transmitted uniformly in all directions, and the expansion takes place only in the direction of the least resistance – upward. The top of the column rises and elevates that part of the earth's crust.

If the uplift of the crust is not accompanied by erosion, the weight of the column of magma and the rocks above it is the same as that of a similar column of the surrounding rocks. Hence the base of the magma column is in equilibrium with the surrounding rocks. Its top, on the other hand, is not: the pressure within it, being hydrostatically transmitted from the bottom, is greater than that in the surrounding rocks. This creates a tendency toward lateral expansion in the upper parts of the column, which, in turn, leads to corresponding deformations of the adjacent strata (Fig. 317*a*). In the more usual case in which the bulge is eliminated by erosion, the magma column and the overlying rocks become lighter than the surrounding mass. Then the pressure of the latter constricts the magma column and squeezes it upward (Fig. 317*b*; note the direction of the arrows at the edges of the column).

In either case the rising magma exerts a strong upward pressure on the earth's crust; the latter, because of its elasticity, resists this pressure.

The magma will continue its upward push until the excess pressure, added to the vertical resistance to deformation, exceeds the resistance to horizontal deformation in the adjacent part of the earth's crust. Beyond that point horizontal deformation follows the uplift, and an orogenic deformation may be initiated.

To illustrate this idea, the authors presented a picture in which the powerful pressure of the rising magma column, on the one hand, and the weight and the elastic resistance of the crust, on the other, crush the softened base of the crustal column, with a lateral squeezing of its material (see Fig. 317, dotted arrows above magma column). In the authors' words, vertical pressure is transformed into horizontal. As already noted, the horizontal current follows an upward slant, along the

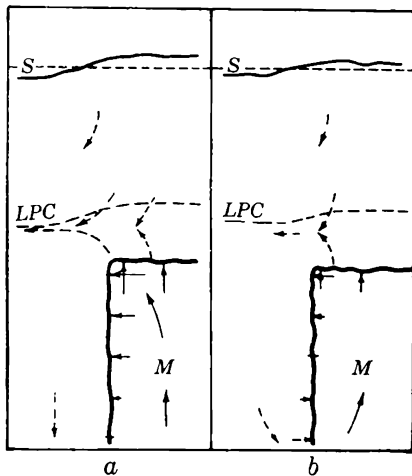


FIG. 317. Diagram of tectogenesis according to the Willis hypothesis: *M*, magma column; *LPC*, level of potential fluidity of rocks; solid arrows, stresses in the column; broken arrows, movement of surrounding rocks; *S*, surface of the earth.

surface of shear. This movement is not symmetrical, since it encounters different resistance in various directions. As a result of this movement, arches of mountains are formed, fringing the central area of depression directly above the magma. The latter, rising ever higher, appears near the surface in the form of large intrusions and, finally, as extrusions. Radioactive elements are carried to the surface with the magma, but since they no longer maintain the high temperature, the magma hardens and a period of quiescence begins, during which new radioactive heat accumulates at depth to initiate another cycle.

As an example of their conceptions, the authors presented a diagram of the structure of the Japanese Islands (see Fig. 316).

THE RADIOMIGRATION HYPOTHESIS

In 1942 to 1943 the geotectonic hypothesis of the present author was published; its main features are the following:

The view that folding is a result of vertical movements of the earth's crust, and of the resulting stresses in the crust, permits this hypothesis to consider only the causes of the oscillatory movements (and of igneous activity). Radioactivity is taken as the chief energy factor. The essential feature of this writer's views, distinguishing them from other hypotheses based on radioactivity, was the concept of an uneven and always changing distribution of the radioactivity within the earth. This premise has met and overcome many difficulties and has apparently paved the way for progress in solving the most general problems of geotectonics.

The change in the distribution of radioactive elements within the earth is connected with the migration of its material. Generally speaking, the redistribution of the radioactive elements is reflected in the formation of large granitic masses in the earth's crust. Granite being the richest among the rocks in radioactive elements, each phase of the formation of granitic magma in the outer mantle of the earth enriches this layer in radioactive elements, undoubtedly at the expense of the deeper layers. Since the formation of a granitic mantle is, generally, an irreversible process, the migration of radioactive elements away from the center is also irreversible. This migration, taking place in phases or pulses along with the formation of granitic intrusions, develops unevenly in both space and time.

The migration of the radioactive elements toward the surface must lead to a general cooling of the subcrustal layers of the earth, as the sources of heat, in their approach to the surface, dissipate their heat into space.

Most likely the formation of granites is related to density stratification of the earth's material, wherein the lighter material (granite) rises while the heavier sinks. Inasmuch as granites have formed throughout the course of geologic history, differentiation of the earth's material has

likewise been occurring throughout geologic time and into the present. Since the earth's substance was previously less differentiated, the radioactive elements rising with the granites must originally have been distributed more uniformly within the earth rather than concentrated chiefly in the crust, as later became the case.

Because of the more widespread distribution of the radioactive elements at depth, the earth in general must have been more intensively heated. It was suggested that this early heating had been so strong as to cause the earth to expand. This expansion, originating deep within the earth, had to overcome the resistance of the solid crystalline crust. After the crust had resisted for a time, cracks appeared in it and the material pushed upward from the depths through the fissures, to merge on the surface as lava flows. This volcanic eruption led to the loss of energy and to cooling of the subcrustal regions. Inasmuch as the eruptions of magma were of different sizes in different places, so did the cooling differ in rate. In places of more rapid cooling, the subcrustal material contracted, and the earth's crust subsided. Thus intrageosynclines appeared, separated by intrageanticlines—zones of less rapid cooling of the subcrustal matter.

It is known from petrology that cooling induces differentiation of the magmatic material. Therefore it is very likely that the differentiation is most intensive under the intrageosynclines. As a result, acidic material, enriched in radioactive elements, accumulates under the axis of the subsiding intrageosyncline, thus building up a local subcrustal focus of heat (Fig. 318). Expansion of the material begins in this focus and is reflected in the central uplift within the intrageosyncline. In the further process of expansion, the persisting subsidence along the margins of the uplift is explained by a horizontal flow of the radioactive acidic material, in the process of differentiation, under the uplift and away from the zones of subsidence, thus causing the zones flanking the central uplift on either side to cool. If the uplift has been sufficiently vigorous, the earth's crust is cracked and the entire cycle is repeated.

Nevertheless, as a result of each cycle, some of the acidic material remains in the earth's crust; thus from cycle to cycle there is an irreversible upward migration of the radioactive elements. This, as we have seen, should have led to a gradual overall cooling of the interior of the earth, thus eliminating the cause of the earth's expansion and thus of the fissuring of its crust and of "magmatic eruptions." In other words, one of the tectonic results of this centrifugal migration of radioactive elements should be a change from extremely active tectonic movements to less active: from geosynclinal to platform conditions.

Since the migration of radioactive elements is uneven—slower in some places and faster in the others—the transition from the original universal geosynclinal conditions to platform conditions has likewise been uneven, occurring first in certain regions and later in others. This has

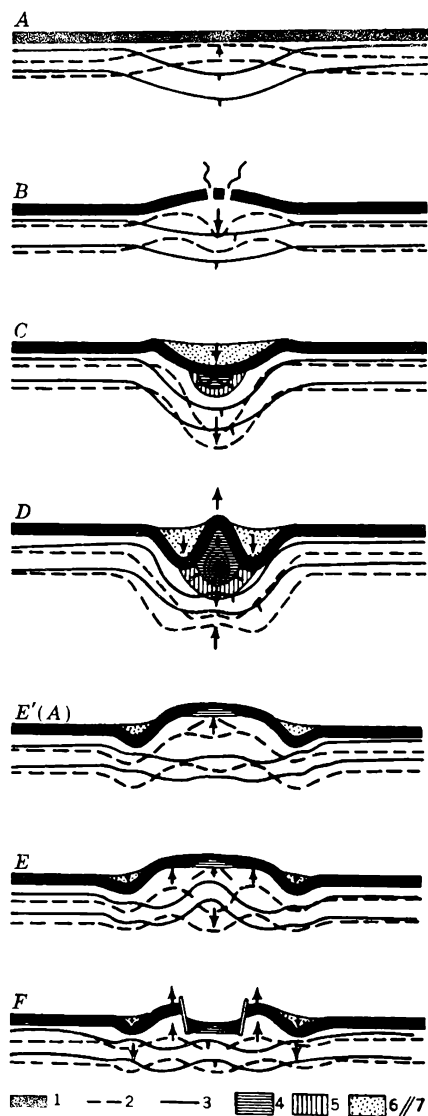


FIG. 318. Development of a geosyncline, according to the radiomigration hypothesis of V. V. Belousov: (1) earth's crust; (2) isotherms; (3) isorads; (4) accumulations of acidic magma; (5) regions of decreased content of radioactive elements; (6) deposits; (7) faults.

resulted in a long-lasting differentiation of the earth's surface into areas in various stages of development. But the general direction of the development from geosynclines to platforms is, in the light of these conceptions, both regular and inevitable.

A dual result of the centrifugal migration of radioactive elements has been noted: both local heating and local cooling are associated with it at various stages of the process. Whenever, in the course of differentiation, radioactive elements accumulate in certain zones beneath the crust, local heating takes place because of the greater concentration of the sources of heat. When the influx of these elements ceases and stationary conditions obtain, the thermal regime tends to become stable through cooling from the original higher temperature. This is determined by the general change in the distribution of radioactive elements toward their greater concentration in the upper layers. The formation of depositional basins and interior depressions is explained by this mechanism. In general, the cooling associated with the centrifugal migration of radioactive elements, according to this hypothesis, leads to the subsidence of large regions as well as to the formation of the ocean basins. As the cooling process develops, the latter grow at the expense of the continents.

It is suggested that the differentiation of subcrustal material does not come to a full stop with the end of the geosynclinal phase but continues at a somewhat slower rate under platform conditions, in which the subsidence of the subgeosyn-

clines is likewise connected with gradual cooling at depth. As a result of differentiation beneath the subgeosynclines, small individual uplifts-discontinuous folds appear on the surface. Thus this hypothesis explains the basic similarity between central uplifts and discontinuous folds that was stressed in Chaps. 19 and 23.

Thus this hypothesis states that the tendency toward thermal equilibrium is the basis for the development of the earth's structure. It has been suggested that originally the earth was an unstable system: the originally widespread dispersion of radioactive elements caused an intensive heating of the interior, which, in turn, led to its expansion. Under these circumstances, the cycle developed until, as a result of differentiation, migration of radioactive elements away from the earth's center began. This, in turn, led to cooling of the depths, contraction of the earth, and almost complete equilibrium throughout the whole system. A connection was established between geotectonic and geochemical processes: the earth's crust is in movement only as long as slow differentiation of the material continues; when the latter nears its end, the movements become quieter.

At the first appearance of this hypothesis, the author himself pointed out some of its shortcomings. Among them is the lack of an explanation for the periodicity and approximate synchronization of the tectonic processes, as well as a certain ambiguity in the inferred relationship between the movements in geosynclines and on platforms.

Enough time has passed since the publication of this hypothesis for its reevaluation in the light of new data. The importance of the centrifugal migration of radioactive elements and of the thermal results of such migration has apparently been confirmed. There is no doubt that the redistribution of radioactive elements with the earth and their uneven dissemination are of great significance in the mechanism of the geotectonic process. On the other hand, it appears that the radiomigration hypothesis underestimated the importance of the force of gravity as an energy factor capable of producing the differentiation of matter, without either magmatic eruptions or local and temporary periods of cooling. The promotion of the force of gravity to the rank of a leading energy factor, upon which the very important but secondary factor of radioactivity is superimposed, opens the door to further advances in understanding the geophysical and geochemical processes at depth. Moreover, it must be said that the original radiomigration hypothesis was too detailed in its description of the processes in the earth's depths. Such a degree of particularization is hardly justified by the present state of knowledge of the physical and chemical properties of the subcrustal material.

CHAPTER 38

CAUSES OF TECTONIC PROCESSES

This concluding chapter will present some observations on the causes of tectonic movements and the nature of the processes in the depths of the planet that now appear the most likely to be correct. These are by no means final, but they are deemed fruitful in charting further progress in an extremely difficult problem (see Belousov, 1951).

DIFFERENTIATION OF MATTER AS A BASIS FOR THE DEVELOPMENT OF THE EARTH

There is sufficient basis for believing that the basic process within the earth that determines the development of the planet is one of stratification of its material, roughly by density. There is no doubt that this process occurs in the crust, and it is apparently absent in the core, as mentioned previously.

Data on certain peculiar aspects of igneous activity have been cited above, particularly on the predominance of basalts among extrusive rocks, which form great thicknesses of lavas remarkably uniform in their composition. Moreover, it is significant that large outpourings of basalt occur in relatively quiet tectonic zones—on platforms. These data strongly suggest that under the earth's crust, composed chiefly of acidic intrusive rocks, lies a thick basalt layer, which feeds these extensive outpourings of basalt. The latter are undoubtedly an emergence on the surface of the material in the earth's depths in its least altered form.

Among the intrusives, on the other hand, acidic rocks predominate, occurring chiefly as batholiths; the mechanism of their formation has been discussed briefly. It was pointed out that the composition of the batholiths depends on the reaction between the original intrusive magmatic material and the intruded rocks. Although basalts penetrate the earth's crust through fissures, undergoing almost no chemical interaction with the host rocks, acidic intrusives are formed under the conditions of a very intensive reaction, taking the form of an assimilation of the intruded rocks, or *granitization*. This, however, does not mean that the primary magmatic material giving rise to the formation of granitic

batholiths was of basaltic composition; it must have been considerably more acidic.

It has been said earlier that along the margins of geosynclines and on platforms, minor intrusions occur that had more a mechanical than a chemical effect on the intruded rocks. The composition of such intrusives apparently has little to do with the surrounding rocks. They are chiefly alkaline, represented by such rocks as syenites, nepheline-syenites, and shonkinites. Most likely such rocks were formed as end products of the differentiation of the basalt in the earth's depths. The fact that such rocks do actually result from differentiation of basalt is shown by the occurrence among the basaltic lavas of the Hawaiian Islands of trachyandesites, trachytes, tephrites, and other alkaline rocks, all of them products of the local differentiation of these lavas (Zavaritskiy, 1944). An even more interesting association is that of the volcanic rocks of Tahiti, where nepheline-syenites, nepheline-monzonites, essexites, and others occur among the differentiation products. Granitization of part of the earth's crust and the formation of granitic massifs probably result from the intrusion of such alkaline magmas into the earth's crust under great pressure and their interaction with sedimentary, metamorphic, and acidic intrusive rocks.

The history of igneous activity reveals that differentiation regularly takes place at depth during the course of a geotectonic stage and under geosynclinal conditions. In geosynclinal areas of subsidence the flows begin with basaltic lavas, followed by intermediate and acidic lavas, the latter predominating toward the end of the first stage. This evolution of the effusive rocks may be interpreted as the result of a differentiation of the basaltic layer, with a gradual accumulation on its surface of more acidic and lighter components. With the formation of the new central uplift, acidic magmas are injected into the earth's crust, which remains in a state of intensive compression; granitic intrusives are formed by granitization of surrounding rocks. Later on, during the stage of fissure intrusions, and at the beginning of the final stage of extrusive activity, magmas of various composition rise and come to the surface, carrying the products of differentiation at depth and themselves being a product of that still active process. Finally, when the supply of differentiation products has been exhausted, a homogeneous primary basalt lava pours out. With each new phase of granitic intrusions, the earth's crust is replenished with new masses of acidic material. The entire sialic mantle consists of multiple granitic intrusions and of the products of their destruction by erosion.

Originally, when geosynclinal conditions prevailed over most of the earth's surface, granitic crust was likewise being formed everywhere. The greater part of the granite crust is consequently very old. The formation of the crust persisted into later epochs, when the platforms appeared

and grew; however, this process has been occurring almost exclusively within the geosynclines, whereas in the areas that subsequently became platforms it is nearly ended. Thus the formation of the upper layer of the earth's crust is a result of differentiation within the outer, basaltic shell of the earth.

It is especially important to understand that differentiation of the matter within the earth did not end with the early stages of its development but it continues up to the present, since the youngest Alpine granitic intrusives in geosynclines and acidic lava flows are all associated with this process.

What are the conditions under which differentiation of the basaltic layer occurs? There is an equilibrium of physicochemical systems, consisting of a number of components and lying in a gravitational field, subject to the laws of thermodynamics. This question will not be dwelt on, except for a reminder that this process depends on the quantitative composition of the system, on the properties of its components, and on the distribution of temperatures and pressures. Any change in a single parameter disturbs the balance and causes a redistribution of the components.

Mechanical equilibrium is one aspect of the more general phenomenon of thermodynamic equilibrium. There is a common tendency to conceive of the processes within the earth as being purely mechanical. Differentiation is often understood exactly in this way: as a purely mechanical separation of the components of the earth according to their respective densities. Such a mechanical process of layering in the earth might be possible if the components of different densities formed bodies sufficiently large that, being plastic, they could flow independently of each other, either rising or sinking according to their respective densities (Lyustikh, 1948a). But such a process cannot occur within a homogeneous solution or a finely dispersed mixture. The difference in behavior between large and small bodies is such that as the body grows, its surface-volume ratio decreases and with it the surface friction. Buoyancy, which is a function of the difference in densities, increases at the same time. Therefore a large body is more easily buoyed up than a small one, the difference in densities being the same.

Moreover, it is difficult to believe that the earth was originally formed of large bodies with different densities. According to O. Yu. Schmidt's theory, the earth must have been formed chiefly of very small particles; its original state, because of the small size of the latter, should rather be imagined as an almost homogeneous mixture of the individual components. It is likely that the layering within the earth developed from such a homogeneous mixture or solution of the components by the separation of acidic material from the basaltic layer. There is no reason to suppose any preexisting large acidic bodies within the basaltic layer.

On the contrary, all the evidence suggests that, except for the specific conditions resulting in differentiation, the basaltic layer is a perfectly homogeneous silicate solution. The formation of layers within such a solution is not a mechanical but rather a physicochemical process, based on molecular diffusion.

It is very important, also, to note that since the redistribution of the components is not mechanical but physicochemical, it should be a reversible process. Thus in those places where differentiation by densities has already taken place, a reverse diffusion of particles might be initiated by an appropriate change in the conditions, and the system would tend toward restoration of its homogeneity.

Finally, for a proper understanding of differentiation, one must keep in mind that it is not atoms that rise and fall within the earth's outer shell but molecules of greater or less complexity, depending on the depth of the process. Therefore some heavy elements, such as the radioactive elements, rise to the surface as constituents of the lighter molecular compounds.

A precise computation of the thermodynamic equilibrium under the conditions existing within the earth is hardly possible as yet. Rough calculations show that an increase in the calcium, magnesium, and iron contents with depth is necessary for a state of equilibrium in the gravitational field of the earth (Ramberg, 1948). The arguments for the occurrence of these changes in composition with depth, not only in the crust but in the mantle of the earth, have already been presented.

One can also have some idea of the direction in which the conditions must change within the system already in equilibrium for the differentiation to be resumed or for the opposite tendency toward homogeneity to appear. Generally speaking, decreases in temperature and pressure promote layering in a gravitational field, whereas a rise in either causes the system to become more homogeneous. This is correct, provided the composition of the system remains unchanged. With simultaneous changes in composition, the conditions are considerably complicated. In a very viscous medium, such as matter deep in the earth, heat possibly assists the differentiation, inasmuch as it increases the mobility of the medium; but this is true only up to a certain limit. Further heating would lead to a redistribution of matter and to greater homogeneity.

Although he concludes that differentiation within the earth is brought about not so much mechanically as physicochemically, the present author in no way denies the significance of mechanical processes. After the physicochemical process has advanced far enough for the material to be separated into large bodies of different densities, some of these bodies might move upward, in a purely mechanical sense, provided the viscosity is sufficiently reduced. This writer thus sees the mechanical

process as a phase and a result of the more general physicochemical process.

It is to be noted, nevertheless, that the conditions of physicochemical and mechanical differentiations are somewhat opposed: though the first, generally speaking, is stimulated by a decrease in temperature—with the possible exception of a very viscous medium in which, within certain limits, heating promotes differentiation—the second is unreservedly facilitated and accelerated by a temperature increase. This contradiction should be considered in any further elaboration of these concepts.

Differentiation is accompanied by complementary phenomena essential to the tectonic mechanism as a whole. These include the thermal results of the migration of radioactive elements: heating and expansion of the material resulting from a concentration of radioactive elements within a certain zone, with its subsequent cooling and contraction upon the establishment of a new thermal equilibrium associated with a greater concentration of radioactive elements near the surface. These thermal phenomena make the tectonic process more intensive: for instance, a "bubble" of acidic material, in its upward rise during mechanical differentiation, is heated by the radioactive elements concentrated within it. As it is heated, it expands, thus exerting a mechanical and physicochemical influence on the surrounding masses and hastening its own upward movement.

The conditions of differentiation must change at depth: first, because of the increase in viscosity with depth as a result of the increasing ambient pressure; second, because of a decrease in the force of gravity, which is known to reach its maximum at about one thousand kilometers, decreasing below that depth to zero at the center of the earth. Both conditions hinder differentiation at great depths.

Differentiation in the deeper parts of the mantle proceeds at a slower pace than in the upper; therein lies a possible explanation for the seismic data pointing to a thinner layering in the crust than in the mantle.

This diversity in the conditions of differentiation at different depths suggests a degree of autonomy at the several levels. This relative autonomy is determined in the first place by the varying rates of differentiation: it is apparently faster in the upper than in the lower levels. Furthermore, differentiation in the upper layers is strongly influenced by surface processes, which have almost no effect at great depths. Finally, a difference in the thermodynamic regime leads to a difference in the conditions of equilibrium; more specifically, the same earth material at different depths is separated into different groups of components.

Among these factors determining the supposed autonomy of the several levels, the difference in the rate of the process is tentatively con-

sidered the most significant. Thus the idea of a differentiation process active at many levels is paramount in this scheme. It will be seen below that this idea involves the possibility of differentiation continuing at certain levels after it has ceased at others. Likewise, it follows that, in the process of differentiation, zones of density inversion may appear within the earth, with heavier material above lighter.

Let us imagine a mantle of originally homogeneous composition. Differentiation occurs throughout its depth, at different rates on different levels. Owing to the more intensive differentiation in a certain upper layer of the mantle, the heavier material accumulates relatively fast at the base of this layer. There it lies above a layer of slower differentiation; the top of this lower layer will have either the average density of this heavier material, or a somewhat lower density if the differentiation in that lower layer, however slow, has caused the lighter material in it to rise.

Such density inversions cannot remain stable. When enough heavier material is accumulated on top of the lighter, the latter may break through, the heavier material sinking while the lighter rises. Thus two layers are involved in an exchange of material; the differentiation becomes more general, and the layering becomes coarser.

These suppositions suggest the concept of a peculiar combination of processes at depth, wherein the establishment of equilibrium (as a result of layering by density) in one level is accompanied by the appearance of unstable zones at greater depths; when such instability leads to a redistribution of the material, the equilibrium on that level is destroyed as well. A process seemingly directed toward the establishment of equilibrium inevitably produces the opposite. We shall attempt to show that this thesis is of far-reaching value in the interpretation of geotectonic processes.

GEOTECTONIC RESULTS OF THE PROCESS OF DIFFERENTIATION WITHIN THE EARTH

The differentiation taking place in the upper part of the mantle is related to the wavelike oscillatory movements of the earth's crust responsible for the development of uplift and subsidence in geosynclines.

A concentrated rise of acidic material, separated and consolidated into a large "bubble," must produce an uplift of the earth's crust above it. This is supported by the fact that uplift, as a rule, is accompanied by outpourings of acidic magma and by intrusive as well as extrusive igneous activity, whereas basic magma is generally associated with areas of subsidence. It is likely that the granite domes so typical of Archean rock series are such bubbles of acidic material risen in the process of differentiation.

This postulated mechanism of the oscillatory movements explains the mutual connection between uplift and subsidence: the rise of a concentrated mass of lighter acidic material involves a parallel sinking of the heavier material. This must be the primary cause of the formation of compensatory downwarps. The exact manner in which differentiation is associated with the wavelike movements of the earth's crust is best seen in the development of a particular geosynclinal subsidence. It has been pointed out that differentiation occurs beneath it, as reflected in the change in composition of the outpouring lavas from basic to acidic. The subsequent formation of a central uplift is related to the accumulation and rise of the acidic differentiation products; the accompanying formation of areas of subsidence is taken to be the result of a compensatory downward movement of the heavier material.

These processes should not be considered within the narrow framework of isostasy. Gravimetric data clearly show some perceptible deviations from isostatic equilibrium existing in nature. The Greater Caucasus, for instance, is overloaded, yet its present rise is proceeding against isostatic forces; on the other hand, the ocean trenches fringing the island arcs are underloaded. These anomalies of isostasy can be partially explained by the fact that the behavior of the components in the physicochemical stage of differentiation is determined not only by their densities but by their much more complex physicochemical properties. Deviations are also produced by the above-mentioned thermal phenomena accompanying the differentiation. Inasmuch as a concentration of radioactive elements in the rising masses of acidic material causes them to become hotter, they must lead to expansion of the material. Expansion intensifies the process of uplift in the earth's crust and the intrusion of acidic masses at the same time that it affects the distribution of densities.

The ensuing cooling associated with the migration of radioactive elements within the earth's crust and with their approach to the surface must lead, on the other hand, to a local shrinking of the material, which cannot fail to affect the gravitational regime. It follows, as already mentioned earlier, that in the most recent concepts, unlike the earlier ones, the most prominent position is occupied by the force of gravity, whereas radioactive heat is still considered important but of secondary significance.

Ye. N. Lyustikh (1948*b*) and V. A. Magnitskiy (1948) have examined the possible causes of vertical movements of the earth's crust in the light of gravimetric data. They concluded that the movements here called wavelike are determined by subcrustal redistribution of material moving from beneath areas of subsidence to beneath uplifts. This, at least, is the picture in the case of platforms; preliminary data, however, suggest that the same is true of geosynclines. These conclusions

agree with the ideas presented here, since differentiation at depth must involve this same redistribution of material.

The transition from geosynclinal to platform conditions may be explained as the result of the cessation of the differentiation process in the upper level. The capacity of matter for differentiation, under any given conditions, is not unlimited but must sooner or later be exhausted. Differentiation ceases when the distribution of the components best suited to the given conditions has been attained. It has already been said that differentiation must be more rapid in the upper layers of the earth than in the lower; hence it follows that its end must occur earlier at shallower than at greater depths.

It was also said (Chap. 16) that the relationship between wavelike movements in geosynclines and platforms suggests a connection between the phenomena occurring in different levels. This may be correlated with the idea suggested here of a multilevel process of differentiation. It seems possible that the intensive wavelike movements of the earth's crust in geosynclines reflect an intensive differentiation within the upper layers of the earth. Platform movements, on the other hand, are determined by differentiation at greater depths. The rapid differentiation within the upper layers terminates earlier, "erasing," so to speak, the geosynclinal movements and leaving only the movements caused by the deeper, slower, and more long-lasting differentiation to be manifested at the surface. This explains the substitution of platform conditions for geosynclinal. The same interpretation explains why platform movements "show through" in geosynclines: the movements at deeper levels are camouflaged by the more intensive movements at the upper levels.

The comparative slowness of the platform movements reflects the corresponding slowness of the deeper differentiation. This idea also explains the fact that the individual areas of uplift and subsidence on platforms are larger than those within geosynclines. Likewise, it has been noted that the redistribution of material at great depth is reflected on the surface by uplifts and depressions chiefly of rounded irregular outlines, as compared with the linear and markedly elongated ovals typical of individual geosynclinal zones. On the other hand, it was pointed out earlier that in the Archean such roundness was also characteristic of the zones of uplift and subsidence produced by differentiation in the upper levels. The appearance of linearity in the disposition of tectonic zones was associated with the formation of systems of primary deep faults. These would, of course, inevitably affect the process of differentiation at depth and produce a corresponding irregularity in it. It is believed that the differentiation must be more intense in the zone adjacent to a deep fault than in other areas, because of movement along the fault and the decrease in pressure in its vicinity.

This discussion of the relationship between tectonic movements and differentiation in the deeper layers refers to the development of uplift and subsidence of the first order (anticlises and synclises). In weaker form the differentiation under a platform may continue in the upper level as well, where it is reflected in smaller structural forms—domical folds, minor arches, etc., often preceded by equivalent local subsidence and accompanied by compensatory depressions—all imitating, on a smaller scale, the central uplifts within geosynclines. This is a weak echo of the differentiation in the upper level, whose violent manifestation marked the geosynclinal stage.

Thus it is suggested that the universally observed complexity of oscillatory crustal movements reflected in the superimposition of movements of different orders is determined by the multilevel nature of the differentiation of the earth's substance. The movement of each order has its own depth of origin; on the surface, however, they are added together and interfere with each other. The interrelationship between the movements of different orders reflects the relationship between their levels and the differences in their distribution and development reflect the degree of autonomy of these levels.

The concept of a leveled differentiation aids in understanding the many peculiarities of the development of oscillatory crustal movements. There are indications that individual areas keep their own persistent tendencies toward uplift or subsidence, regardless of the passage of waves of subsidence or uplift over them. Thus if an area has a tendency toward uplift, a passing wave of subsidence (as, for instance, in the migration of a foredeep) would here induce a subsidence less pronounced than in the neighboring areas lacking such tendencies toward uplift. On this basis, M. V. Gzovskiy (1948a) introduced his concepts of elementary geotectonic regions. This also applied to N. S. Shatskiy's suggestion (1945) that there is a narrowing and complete wedging out of the foredeeps, where they are superimposed on intensively rising platform margins. All such phenomena are explicable only by the fact that different oscillatory movements originate at different depths and are literally superimposed on each other in their passage through the different levels and in their combination on the surface.

The possibility of a disruption of the equilibrium on the border between two levels suggests an explanation for the activation of a platform. A break-through of earth material in a zone of "density inversion" (after equilibrium in the upper level has been attained and platform conditions established) leads to a revival of differentiation and of intensive oscillatory tectonic movements along with it. Since here the material of both the upper and the lower layers is drawn into the circulation, there is a possible explanation for the difference between the structures of the earth's crust beneath young geosynclines and activated

platforms: it has been pointed out that the uplifts in geosynclines have granitic roots, whereas basaltic roots have been discovered under the uplifts on activated platforms.

At the present time no attempt is made to answer the question of the magnitude of the several levels nor that of their structure and composition. In any event, the upper level must embrace all the earth's crust with its granitic and basaltic layers, in addition to another and deeper layer, so that its total thickness can hardly be less than 100 km.

The problem of general oscillations of various orders is a complex one. This writer believes it is significant that the epochs of general continental uplift coincide with the wavelike individual uplifts. Inasmuch as the wavelike movements are associated with differentiation and with the rise of lighter material, it may be conjectured that the general oscillations are induced by a periodic change in the temperature of the earth's substance: a rise in temperature increases the volume of the material, but at the same time it decreases its viscosity and facilitates its mechanical differentiation. Therefore, against the background of a general uplift of large areas of the earth's crust caused by the expansion of the material, there is an intensification of the rise of acidic material and of wavelike movements of any order. The reasons for the suggested fluctuations in the temperature are not known.

The process of formation of the oceans and continents still involves much that is puzzling. The suggestion was made earlier that certain extensive regions of the earth's crust at some stage of their development undergo basification and a sharp change in their general composition and that the resulting increase in their weight leads to their subsidence and the emergence of water from the depths. This transformation of a granitic-basaltic crust into a water-basaltic crust is a radical alteration in the general tendency of development of the earth's crust. The significance of these changes is still completely unknown. It is also unknown how such intensive antithetical recent processes as the formation of ocean basins and the activation of platforms on continents are related to each other.

Very recent data indicate the "oceanic" nature of a number of seas, such as the Caribbean, the Mediterranean, the Black, the southern part of the Caspian, and the Gulf of Mexico. Is it not possible that seas of this type represent the initial stage in the formation of ocean basins?

A curious analogy suggests itself between the theory of the formation of the oceans on the earth as described here and the history of the "seas" on the moon. A. V. Khabakov (1950) has shown convincingly that the moon's "seas" were formed as the result of subsidence, along fault lines, of parts of the moon's surface that earlier had a high "continental" relief. These elevated parts of the lunar surface are characterized by a special type of relief whose main feature is the develop-

ment of numerous craters of volcanic origin. The bottoms of the "seas" are flat, and have visual properties strongly suggesting that they are covered by basic lava flows. The faulted margins of the "seas" unconformably intersect the structural elements of the "continents," thus indicating a formation of "marine" basins by partial subsidence of "continental" massifs. A still more convincing piece of evidence is the vestiges of "continental" relief, which are half-inundated by the basalt flows on the bottom of the "seas" and can be seen through these flows everywhere, with varying degrees of clarity. This, however, is as far as the analogy between the earth and the moon can be drawn.

SOME CONCLUSIONS

It follows from what has been said above that the tectonic process is determined by an intricate complex of phenomena, based on an intermittent differentiation of the earth's substance. This is a multilevel and seemingly contradictory process: after a local equilibrium has been attained in the outer level of the earth and the related tectonic movements have died down, the differentiation and the stronger tectonic movements may be revived. In general, however, the tendency of the process is toward an increasingly thinner layering of the material within the earth. It is possible that periodic changes in the temperatures at depth are superimposed upon the process of differentiation, thereby inducing a corresponding periodic acceleration and retardation of the differentiation.

Differentiation is connected with the redistribution of radioactive elements, which, in turn, leads to a difference in temperature: to the heating of some parts of the earth's interior and relative cooling of others. Expansion and compression associated with the temperature changes add to the intensity of the uplifts and cause isostatic anomalies. Finally, it is possible that to all this must be added the chronologically more recent process—whose course runs in the opposite direction—of local increase in the basic components of the earth's crust at the expense of the acidic, with the resulting formation of ocean basins.

This is the present author's hypothesis, which, of course, makes no claims to infallibility; nevertheless, it seems to indicate the proper direction for further research. Investigations should be conducted along the following lines:

In the field of geotectonics, there should be:

1. A study of the vertical movements of the earth's crust, wavelike as well as general oscillatory movements. Such a study should produce the most complete data on the history of the vertical crustal movements over the entire surface of the continents and in the oceans, over the entire course of geologic history, and for the entire range of movements.

2. A study of all the manifestations of igneous activity, to clarify the connection between the composition of magma and the occurrence of igneous rocks, on one hand, and the movements of the earth's crust, on the other; and also to determine the history of igneous activity for as long a time in the past as possible.

In the field of geophysics, there should be:

1. A refinement of seismic data on the discontinuities between the layers within the earth and a study of the structural surface relief of these discontinuities. Assuming that the course of differentiation at depth is uneven, that relief should be complex. It is also anticipated that the number and distribution of the discontinuities will differ in places with different structures and histories. This should include a search for zones of density inversion.

2. A study of the earth's gravitational field in the light of the structure and development of the earth's crust as revealed by geologic methods. In conjunction with the seismic data, such a study should clarify the redistribution of the masses within the earth's crust and provide information on the history of formation of the crust.

3. A study of the earth's magnetic field, to determine its connection with the movements and the major structures of the earth's crust.

4. A study of the changes in the amount of water on the earth.

In the field of thermodynamics of the earth, there should be:

1. An investigation, using both experiments and computations, of the physical and chemical properties of substances under the conditions of pressures and temperatures prevailing at various depths within the earth.

2. An experimental and analytical study of conditions causing the loss of thermodynamic equilibrium at various depths within the earth; of conditions determining the separation of the components, their diffusion, and other elements in the mechanism of differentiation.

3. An analytical study of the thermal field of the earth, in connection with the radioactive origin of heat and with the changing distribution of the heat sources resulting from the differentiation and the gradual concentration of its acidic products, enriched with radioactive elements, in the outer layers of the earth.

BIBLIOGRAPHY

1. Adams, F. A., *The Birth and Development of the Geological Sciences*, Baltimore, 1938.
2. "A Discussion of the Floor of the Atlantic Ocean," *Proc. Roy. Soc. (London)*, ser. A, no. 1150, vol. 222, 1954.
3. Afanas'yev, G. D., "Opyt sopostavleniya intruzivnykh kompleksov nekotorykh oblastey S.S.S.R. (A comparison of intrusive complexes in certain regions of the U.S.S.R.)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 4, 1949.
4. Ahrens, L. H., K. Rankama, and S. K. Runcorn (eds.), *Physics and Chemistry of the Earth*, vol. I, London, 1956.
5. Airy, G. B., "On the Computation of the Effect of the Attraction of Mountain Masses," *Phil. Trans. Roy. Soc. (London)*, vol. 145, pp. 101-104, 1855.
6. Ampferer, O., "Über das Bewegungsbildung von Faltengebirgen," *J. d. K. K. Geol. Reichsanstalt (Wien)*, bd. 56, 1906.
7. Anderson, E. M., *The Dynamics of Faulting and Dyke Formation with Applications to Britain*, 1942.
8. Anderson, V., *Geology in the Coastal Atlas of Western Algeria*, G. S. A. Memoir 4, 1936.
9. Andree, K., *Über die Bedingungen der Gebirgsbildung*, 1914.
10. Argand, E., "L'Arc des alpes occidentales," *Ecl. Geol. Helv.*, bd. 14, 1916.
11. Aristotle, *Météorologie*, in *Oeuvres d'Aristote*, Paris, 1863.
12. Arkhangel'skiy, A. D., *Geologicheskoye stroyeniye i geologicheskaya istoriya S.S.S.R. (The Geologic Structure and Geologic History of the U.S.S.R.)*, Gosgeolizdat, 1941.
13. Arkhangel'skiy, A. D., *Geologicheskoye stroyeniye i geologicheskaya istoriya S.S.S.R. (The Geologic Structure and Geologic History of the U.S.S.R.)*, 4th ed., vols. I and II, Gosgeolizdat, 1947-1948.
14. Arkhangel'skiy, A. D., and N. M. Strakhov, *Geologicheskoye stroyeniye i istoriya razvitiya chernogo morya (The Geologic Structure and the History of the Black Sea)*, Izd. A. N. S.S.S.R., 1938.
15. Arkhangel'skiy, A. D., *Geologicheskoye stroyeniye S.S.S.R.: Yevropeyskaya i sredne-aziatskaya chasti (The Geologic Structure of the U.S.S.R.: The European and Central Asian Parts of the Soviet Union)*, Geol.-Razv. Izd., 1932.
16. Arkhangel'skiy, A. D., *Geologicheskoye Stroyeniye S.S.S.R.: Zapadnaya chast' (The Geologic Structure of the U.S.S.R.: The Western Part)*, Onti, 1934.
17. Arkhangel'skiy, A. D., et al., "Geologicheskoye znachenie anomalii sily tyazhesti v S.S.S.R. (The Geologic Significance of Gravitational Anomalies in the U.S.S.R.)," *Izv. A. N. S.S.S.R., ser. Geol.*, no 4, 1937.
18. Arkhangel'skiy, A. D., et al., *Kratkiy ocherk geologicheskoy struktury i geologicheskoy istorii S.S.S.R. (Brief Outline of the Geologic Structure and Geologic History of the U.S.S.R.)*, Izd., A. N. S.S.S.R., 1937.
19. Arkhangel'skiy, A. D., "O nekotorykh spornykh voprosakh tektonicheskoy terminologii i tektoniki S.S.S.R. (Some Controversial Problems of Tectonic Terminology and of the Tectonics of the U.S.S.R.)," *Izv. A. N. S.S.S.R., ser. Geol.*, No. 1, 1939.

20. Arkhangel'skiy, A. D., et al., "Tektonika dokembriyskogo fundamenta vostochnoyevropeyskoy platformy po dannym obshchey magnitnoy s'emki S.S.S.R. (The Tectonics of the Precambrian Basement of the Eastern European Platform, Based on a Magnetic Survey of the U.S.S.R.)," *Izv. A. N. S.S.S.R., ser. Geogr. i Geof.*, No. 2, 1937.
21. Arkhangel'skiy, A. D., *Vvedeniye v izucheniye geologii evropeyskoy rossii (Introduction to the Study of the Geology of European Russia)*, Giz, 1923.
22. Arni, P., "Tektonische Grundzuege Ostanatoliens und benachbarter Gebiete," *Veroeff. des. Inst. fuer Lagerstaettenforschung der Tuerkei*, ser. B., Abh. 4, 1939.
23. Aschauer, H., "Die östliche Endigung des Appenins und ihre paläogeographische Entwicklung," *Abh. Ges. Wiss. zu Goettingen, Math.-Phys. Kl.*, folge III, 15, 1936.
24. Avrov, V. Ya., and M. I. Barenboym, "Nekotoryye novyye fakticheskiye dannyye o glubinnom stroyenii yuzhnoy emby (Some New Data on the Deep Structure of the Southern Emba Area)," *DAN*, vol. 78, no. 6, 1951.
25. Azhgirey, G. D., "Uchastiye drevnego kristallicheskogo fundamenta v al'piyskoy skladchatosti tsentral'nogo kavkaza (The Part Played by the Ancient Crystalline Basement in the Alpine Folding of Central Kazakhstan)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 26, no. 4, 1951.
26. Balk, R., *Structural Behavior of Igneous Rocks*, 1937.
27. Balsley, J. R., "Deformation of Marble under Tension and High Pressure," *Trans. Am. Geophys. Un., Nat. Res. Council*, pt. II, 1941.
28. Bärtling, R., "Das Verhältnis zwischen Sedimentation und Tektonik im Ruhrbezirk," *C. R. II, Congrès pour l'avancement des études de stratigraphie carbonifère*, Heerlen, 1927.
29. Barrell, J., "Rhythms and the Measurement of Geologic Time," *G. S. A. Bull.*, vol. 28, pp. 745-904, 1917.
30. Barton, D. C., "The American Salt Dome Problems in the Light of the Rumanian and German Salt Domes," *Bull. A. A. P. G.*, vol. 9, no. 9, 1925.
31. Bauman, V. N., "K voprosu o sbrosakh, sdvigakh i drugikh smeshcheniyakh zhil i plastov (Normal Faults, Strike-slip Faults, and Other Displacements of Veins and Beds)," *Zap. Gorn. In-ta*, vol. I, no. 1, 1907.
32. Behrmann, R. B., "Die Faltenboegen des Appenins und ihre palaeogeographische Entwicklung," *Abh. Ges. Wiss. zu Goettingen, Math.-phys. Kl.*, folge III, h. 15, 1936.
33. Belitskiy, A., "K voprosu o mekhanizme obrazovaniya klivazhnykh treshchin (The Mechanics of the Formation of Cleavage Fractures)," *Tr. Gorno-Geol. In-ta Zap.-Sib. Filiala A. N. S.S.S.R.*, vyp. 6, Novosibirsk, 1949.
34. Beloussov, V. V., "Bol'shoy kavkaz: Ch. I-III (The Greater Caucasus: Parts I-III)," *Tr. Tsniigri*, vyp. 108, 121, 126, 1938-1940.
35. Beloussov, V. V., "G. B. Sossyur: Pervyy issledovatel' stroyeniya al'p (J. B. Saussure: First Investigator of the Structure of the Alps)," *Priroda*, no. 1, 1940.
36. Beloussov, V. V., "Gipotezy podnyatiya i kontraktsii v geotektonike (The Hypotheses of Uplift and Contraction in Geotectonics)," *Priroda*, no. 1, 1939.

37. Belousov, V. V., *Kolebatel'nye dvizheniya zemnoy kory, ikh razvitiye, svoystva i zadachi ikh izucheniya: Tr. soveshch. po metodike izucheniya dvizheniy i deformatsiy zemnoy kory (The Development and the Characteristics of Oscillatory Crustal Movements and Their Further Study: Transactions of the Conference on Methods of Studying the Movements and Deformation of the Earth's Crust)*, Geodezizdat, 1948.
38. Belousov, V. V., I. V. Kirillova, and A. A. Sorskiy, "Kratkiy obzor seysmichnosti kavkaza v sopostavlenii s ego tektonicheskim (A Brief Survey of Seismic Activity in the Caucasus in the Light of Its Tectonic Structure)," *Izv. A. N. S.S.S.R., ser. Geof.*, no. 5, 1952.
39. Belousov, V. V., "K voprosu o mekhanizme obrazovaniya orientirovki mineralov v gornykh porodakh (The Mechanism Involved in the Orientation of the Minerals in Rocks)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 25, no. 5, 1950.
40. Belousov, V. V., "Migratsiya radioelementov i razvitiye struktury zemli: I-II (The Migration of Radioactive Elements and the Development of the Earth's Structure: I-II)," *Izv. A. N. S.S.S.R., ser. Geogr. i Geof.*, no. 6, 1942; no. 3, 1943.
41. Belousov, V. V., "Moshchnost' otlozheniy kak vyrazhenie rezhima kolebatel'nykh dvizheniy zemnoy kory (The Thickness of Deposits as a Reflection of the Oscillatory Movements of the Earth's Crust)," *Sov. Geol.*, no. 2-3, 1940.
42. Belousov, V. V., "Obshchiye zakonomernosti geotektonicheskogo protsessa (General Laws Governing the Geotectonic Process)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 5, 1948.
43. Belousov, V. V., "O nekotorykh rezul'tatakh i perspektivakh geoktonofizicheskikh issledovaniy (Some Results and Prospects of Geotectonophysical Investigations)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 2, 1958.
44. Belousov, V. V., "O geologicheskoy stroynii okeanov (Geologic Structure of the Oceans)," *Priroda*, no. 5-6, 1942.
45. Belousov, V. V., "O kolebatel'nykh dvizheniyakh zemnoy kory (On the Oscillatory Movements of the Earth's Crust)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 2, 1938a.
46. Belousov, V. V., "O proiskhozhdenii skladchatosti (The Origin of Folding)," *Sov. Geol.*, vol. 16, 1947.
47. Belousov, V. V., "Osnovnyye cherty tektoniki tsentral'nogo i yuzhnogo kitaya (Main Features of the Tectonics of Central and Southern China)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 8, 1956.
- 47a. Belousov, V. V., "Osnovnyye voprosy mekhanizma skladkoobrazovaniya (Basic Problems of the Mechanics of Folding)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 22, no. 3, 1947.
48. Belousov, V. V., "P. S. Pallas: Geolog i puteshestvennik (P. S. Pallas: Geologist and Explorer)," *Priroda*, no. 3, 1941.
49. Belousov, V. V., "Posloynnoye pereraspredeleniye materiala v zemnoy kore i skladkoobrazovaniye (Redistribution of the Material within the Beds in the Earth's Crust and Its Bearing on Folding)," *Sov. Geol.*, vyp. 39, 1949.
50. Belousov, V. V., "Problemy vnutrennego stroeniya zemli i yeye razvitiya: Ch. I-II (The Internal Structure of the Earth and Its Development: Parts I-II)," *Izv. A. N. S.S.S.R., ser. Geof.*, no. 1-2, 1951.

51. Belousov, V. V., "Rol' vremeni v geologicheskikh protsessakh (The Role of Time in Geologic Processes)," *Priroda*, no. 1-2, 1942.
52. Belousov, V. V., "Skhema tektoniki zemli (An Outline of the Earth's Tectonic Structure)," *DAN*, vol. 68, no. 1, 1949.
53. Belousov, V. V., "Tektonicheskiye nablyudeniya vo frantsuzskikh al'pakh v 1955 g. (Tectonic Observations in the French Alps in 1955)," *Sov. Geol.*, vol. 54, 1956.
54. Belousov, V. V., and M. V. Gzovskiy, "Tektonicheskiye usloviya i mekhanizm vozniknoveniya zemletryaseniy (The Tectonic Conditions and Mechanism of Earthquake Origin)," *Trudy Geofiz. In-ta A. N. S.S.S.R.*, no. 25, p. 152, 1954.
55. Belousov, V. V., "Teoriya zemli' Dzhemsa Gettona (James Hutton's Theory of the Earth)," *Priroda*, no. 7-8, 1938b.
56. Belousov, V. V., "Tipy i proiskhozhdeniye skladchatosti (Types and Origin of Folding)," *Sov. Geol.*, no. 1, 1958.
57. Belousov, V. V., and M. V. Gzovskiy, "Geosinklinali, ikh stroeniye, istoriya, i zakony razvitiya: I. Kaledonskaya geosinklinal' velikobritanii (The Structure, History, and Laws of Development of Geosynclines: I. The Caledonian Geosyncline of Great Britain)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 20, no. 5-6, 1945.
58. Belousov, V. V., M. V. Gzovskiy, and A. B. Goryachev, "O strukture vostochnykh al'p v svyazi s nekotorymi obshchimi tektonicheskimi predstavleniyami (The Structure of the Eastern Alps from the Standpoint of Certain General Tectonic Concepts)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 26, no. 1-2, 1951.
59. Belousov, V. V., and K. I. Kuznetsova, "K voprosu o fizicheskikh usloviyakh obrazovaniya tektonicheskikh razryvov (The Physical Conditions Governing the Formation of Tectonic Faults)," *Izv. A. N. S.S.S.R., ser. Geogr. i Geof.*, vol. 8, no. 6, 1949.
60. Belousoff, W. W., "Grundfragen der allgemeinen Geotektonik," *Geol. Rundschau*, bd. 45, h. 2, 1956.
61. Belousoff, W. W., "Einige allgemeine Fragen der Tektonik an der Nahtstelle zwischen Krim und Kaukasus," *Geologie*, jahrg. 7, h. 3-6, 1958.
62. Belyayevskiy, N. A., "Al'piyskaya tektonika zapadnogo kuen'lunya (The Alpine Tectonics of Western Kuen Lun)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 2, 1949.
63. Bertrand, M., "La Chaîne des Alpes et la formation du continent européen," *Bull. Soc. géol. fr., Sér. III*, vol. 15, pp. 423-447, 1886-1887.
64. Betz, F., and H. H. Hess, "The Floor of the North Pacific Ocean," *Geogr. Rev.*, January, 1942.
65. Bilibin, Yu. A., "Voprosy metallogenicheskoy evolyutsii geosinklinal'nykh zon (Problems of the Metallogenic Evolution of Geosynclinal Zones)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 4, 1948.
66. Billings, M. P., *Structural Geology*, 1946.
67. Birch, F., J. F. Schairer, and H. C. Spicer, *Spravochnik dlya geologov po fizicheskim konstantam (A Handbook of Physical Constants)*, Izd-vo Inostr. Lit., 1949.
68. Bogdanov, A. A., "Nesoglasiya, ikh tipy i znachenie ikh izucheniya (Unconformities, Their Types and the Importance of Studying Them)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 2, 1949a.

69. Bogdanov, A. A., "Osnovnyye cherty tektoniki vostochnykh karpát (Main Features of the Tectonics of the Eastern Carpathians)," *Sov. Geol.*, vol. 40, 1949b.
70. Bogdanov, A. A., "Zemlyetryaseniya v severnykh tsepyakh tyan'-shanya 1911 g (Earthquakes in the Northern Ranges of the Tien Shan in the Year 1911)," *Tr. Geol. Kom.*, nov. ser., vyp. 89, 1914.
71. Bogdanov, A. A., *Tektonika ishimbayevskogo priural'ya (The Tectonic Structure of the Urals Region in the Vicinity of Ishimbayevo)*, Izd. MOIP, 1947.
72. Bogdanov, A. A., "Zavisimost' intensivnosti klivazha ot moshchnosti plasta (The Intensity of Cleavage as Related to the Thickness of the Bed)," *Sov. Geol.*, vol. 16, 1947.
- 72a. Bogdanovich, K. I., et al., "Zemletryaseniya v severnykh tsepyakh tyan'-shanya, 1911 g (Earthquakes in the Northern Tien Shan Ranges in 1911)," *Tr. Geol. Kom.*, nov. ser., vyp. 89, 1914.
73. Borisov, A. A., and N. I. Buyalov, "K mekhanike obrazovaniya grabenov embenskikh kupolov (The Mechanics of the Formation of Grabens in the Emba Domes)," *Neft. Khoz.*, no. 5, 1938.
74. Borisyak, A. A., "Teoriya Geosinklinal'ey (The Theory of Geosynclines)," *Izv. Geol. Kom.*, vol. 10, 1924.
75. Born, A., *Der geologische Aufbau der Erde: Handbuch der Geophysik*, 1932.
76. Brinkmann, R., "Betikum und Keltiberikum in Suedostspanien," *Abh. ges. Wiss. zu Goettingen, Math.-phys. Kl.*, folge III, h. 1, 1931.
77. Brod, I. O., "O klassifikatsii neftyanykh zalezhey po ikh formam (On the Classification of Oil Deposits by Their Structures)," *Mat-ly Geol. Kongr.*, *Tr. XVII sessii*, vol. 4, 1940.
78. Bronguleyev, V. V., "Amagmaticheskies in'yektsionnyye yavleniya na platforme (Nonmagmatic injections on Platforms)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 12, no. 6, 1947.
79. Bronguleyev, V. V., "Melkaya skladchatost' platformy: Mat-ly k pozn. geol. stroyeniya S.S.S.R. (Small-scale Folding on Platforms: Materials on the Geologic Structure of the U.S.S.R.)," *MOIP*, nov. ser., vyp. 14/18, 1951.
80. Bronguleyev, V. V., "O printsipe klassifikatsii skladchatykh struktur zemnoy kory (Principles Underlying the Classification of Structures in the Crust)," *Izv. A. N. S.S.S.R.*, ser. Geol., no. 1, 1949.
81. Bubnoff, S., *Geologie von Europa*, bd. I, 1926; bd. II, t. 1, 1930; bd. II, t. 2, 1930.
82. Bubnoff, S., *Geologie von Europa*, bd. II, t. 3, 1936.
83. Bubnoff, S., *Geologiya yevropy (The Geology of Europe)*, ONTI, 1935, vol. 2.
84. Bucher, W. H., *The Deformation of the Earth's Crust*, 1933.
85. Bucher, W. H., "Submarine Valleys and Related Geologic Problems of the North Atlantic," *G. S. A. Bull.*, vol. 51, no. 4, 1940.
86. Bucher, W. H., "Versuch einer Analyse der grossen Bewegungen der Erdkruste," *Geol. Rundschau*, h. 34, 1939.
87. Burkar, Zh. (Jacques Bourcart), *Rel'ef okeanov i morey (Relief of the Oceans and Seas)*, Izd-vo In. Lit., 1953.

88. Carr, D. R., and J. L. Kulp, "Age of a Mid-Atlantic Ridge Basalt Boulder," *G. S. A. Bull.*, vol. 64, no. 2, 1953.
89. Chardin, T., and F. Licent, "On the Geology of the Borders of the Ordos," *Bull. Soc. Geol. China*, vol. 3, 1924.
90. Cherskiy, I. D., "K geologii vnutrenney azii (On the Geology of the Interior Regions of Asia)," *Tr. SPB Ob-va Yestestvoisp.*, XIII, vyp. 2, 1886.
91. Chertkova, Ye. I., "Nekotoryye rezul'taty modelirovaniya tektonicheskikh razryvov (Some Results of the Modeling of Tectonic Faults)," *Izv. A. N. S.S.S.R., ser. Geogr. i Geof.*, t. 14, vyp. 5, 1950.
92. Chubb, L. G., "The Structure of the Pacific Basin," *Geol. Mag.*, vol. 71, no. 7, 1934.
93. Clark, B. L., "Tectonics of the Coast Ranges of Middle California," *G. S. A. Bull.*, vol. 41, no. 4, 1930.
94. Cloos, E., *Lineation: A Critical Review and Annotated Bibliography*, G. S. A. Memoir 18, 1946.
95. Cloos, E., "Öolite Deformation in the South Mountain Fold, Maryland," *G. S. A. Bull.*, vol. 58, no. 9, 1947.
96. Cloos, H., "Hebung-Spaltung-Vulkanismus," *Geol. Rundschau*, bd. 30, h. 4A, 1939.
97. Crowell, J. C., "Submarine Canyons Bordering Central and Southern California," *J. Geol.*, vol. 60, no. 1, 1952.
98. Daly, R. A., *Igneous Rocks and the Depths of the Earth*, 1933.
99. Dana, J. D., *Corals and Coral Islands*, London, 1875.
100. Dana, J., "On the Origin of Mountains," *Am. J. Sci.*, 3d ser., vol. 5, 1873; "On Some Results of the Earth's Contraction from Cooling," *Am. J. Sci.*, 3d ser., vol. 5, pp. 423-443; vol. 6, 1873.
101. Dementitskaya, R. M., "O vozmozhnoy zavisimosti mezhdu tolshchinoy zemnoy kory i vosrastom skladchatosti (A Possible Correlation between the Thickness of the Earth's Crust and the Age of Folding)," *Sov. Geol.*, no. 6, 1958.
102. De Sitter, L. U., *Structural Geology*, 1956.
103. De Terra, H., "Himalayan and Alpine Orogenies," *Int. Geol. Congr.*, XVI sess., vol. 2, 1936.
104. Dietz, R. S., and H. W. Menard, "The Hawaiian Swell, Deep and Arch and the Subsidence of the Hawaiian Islands," *J. Geol.*, vol. 61, no. 2, 1953.
105. Dobrin, M. B., "Some Quantitative Experiments on a Fluid Salt Dome Model and Their Geological Implications," *Am. Geoph. Union, 22d Ann. Meeting, Section of Tectonophysics: Reports and Papers*, 1941.
106. Dobrin, M. B., Beauregard Parkins, and B. L. Snavelly, "Subsurface Constitution of Bikini Atoll as Indicated by a Seismic Refraction Survey," *G. S. A. Bull.*, vol. 60, no. 5, 1949.
107. Dominov, Ye. A., "Yurskiye depressii vysokogornogo kavkaza (The Jurassic Depressions in the High Caucasus Mountains)," *Sov. Geol.*, no. 11, 1958.
108. Dumitrashko, N. V., "Molodost' i drevnost' rel'yefa yugo-vostochnoy sibir: Problemy geomorfologii (The Recent and Ancient Relief of

- Southeastern Siberia: Problems of Geomorphology)," A. N. S.S.S.R., *Tr. In-ta Geogr.*, vyp. 39, 1948.
109. Dutton, C. E., "On Some of the Greater Problems of Physical Geology," *Bull. Phil. Soc., Wash.*, vol. 11, 1892.
 110. Eardley, A. J., *Structural Geology of North America*, 1951.
 111. Eardley, A. J., "Tectonic Relations of North and South America," *Bull. Am. Ass. Petr. Geol.*, vol. 38, no. 5, 1954.
 112. Emery, K. O., "Submarine Geology of Bikini Atoll," *G. S. A. Bull.*, vol. 59, 1948.
 113. Ewing, M., G. P. Woollard, and A. C. Vine, "Geophysical Investigations in the Emerged and Submerged Atlantic Coastal Plain: Part IV," *G. S. A. Bull.*, vol. 51, no. 12, pt. 1, 1940.
 114. Ewing, M., I. L. Worzel, N. C. Steenland, and F. Press, "Geophysical Investigations in the Emerged and Submerged Atlantic Coastal Plain: Part V," *G. S. A. Bull.*, vol. 61, no. 9, 1950.
 115. Ewing, M., G. P. Woollard, A. C. Vine, and I. L. Worzel, "Recent Results in Submarine Geophysics," *G. S. A. Bull.*, vol. 57, no. 10, 1946.
 116. Ewing, M., and F. Press, "Structure of the Earth's Crust," *Handbuch der Physik*, bd. 47, Berlin, 1955.
 117. Fairbairn, H. W., *Strukturnaya petrologiya deformirovannykh gornyykh porod (The Structural Petrology of Deformed Rocks)*, Izd. Inostr. Lit., 1949.
 118. Fersman, A. Ye., *Geokhimiya (Geochemistry)*, ONTI, 1933, vol. 1.
 119. Filonenko-Borodich, M. I., et al., *Kurs soprotyvleniya materialov (A Course in the Strength of Materials)*, vol. I, Gostekhizdat, 1949.
 120. Fotiadi, E. E., "K voprosu stroyeniya dokembriyskogo skladchatogo osnovaniya russkoy platformy (On the Structure of the Pre-Cambrian Folded Basement of the Russian Platform)," *DAN*, vol. 57, no. 8, 1947.
 121. Fourmarier, P., "Les Traits directeurs de l'évolution géologique du continent africain," *C. R. Congrès géol. int. XIV sess.*, fasc. III, 1928.
 122. Fridman, Ya. B., *Mekhanicheskiye svoystva metallov (The Mechanical Properties of Metals)*, Oborongiz, 1946.
 123. Gerodot, *Istoriya v devyati knigakh (Herodotus, History, Ten Books)*, 1888.
 124. Gerth, H., *Geologie Südamerikas*, vols. I-II, 1932-1935.
 125. Gerth, H., *Der geologische Bau der südamerikanischen Kordillere*, 1955.
 126. Geszti, J., "Schichtungsvorgang in einem inhomogenen schweren weltkörper hoher Temperatur," *Gerlands Beiträge zur Geophysik*, bd. 49, h. 1/2, 1937.
 127. Gignoux, M., "La Tectonique d'écoulement par gravité et la structure des alpes," *Bull. Soc. géol. France*, vol. 18, no. 8-9, 1948.
 128. Gilbert, G. K., *Lake Bonneville*, U.S.G.S. Monograph 1, 1890.
 129. Glangaud, L., "La Rôle du socle dans la tectonique du jura," *Ann. Soc. géol. belgique*, t. 73, 1949-1950.
 130. Goldschmidt, V. M., "Stammestypen der Eruptivgesteine," *Vidensk Skrifter, I., Math-Naturw. Kl.*, no. 10, 1922.
 131. Goldschmidt, V. M., "Über die Massenverteilung im Erdinnern," *Naturwissenschaften*, h. 42, 1922.

132. Golovkinskiy, N. A., *Materialy dlya geologii rossii (Materials on the Geology of Russia)*, vol. I, 1869.
133. Goranson, R. W., "The Melting of Granite," *Am. J. Sci.*, vol. 22, p. 481, 1931; vol. 23, p. 227, 1932.
134. Gorshkov, G. P., and A. Ya. Levitskaya, "Nekotoryye dannyye po Seysmotektonike kryma (Some Data on the Seismic Activity and Tectonics of the Crimea)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 22, no. 3, 1947.
135. Grabau, A., *Stratigraphy of China*, pt. I, 1923–1924; pt. II, 1923.
136. Gregory, J. W., "The Geological History of the Pacific Ocean," *Quart. J. Geol. Soc. (London)*, vol. 86, 1930.
137. Griggs, D. T., "Deformation of Rocks under High Confining Pressures," *J. Geol.*, vol. 44, no. 5, 1936.
138. Griggs, D. T., "Experimental Flow of Rocks under Conditions Favoring Recrystallization," *G. S. A. Bull.*, vol. 51, no. 7, 1940a.
139. Griggs, D. T., "Deformation of Rocks in the Laboratory," *Trans. Am. Geoph. Un., Nat. Res. Council*, pt. II, 1940b.
140. Griggs, D. T., "A Theory of Mountain Building," *Am. J. Sci.*, vol. 237, 1939.
141. Grout, F. G., "Scale Models of Structures Related to Batholiths," *Am. J. Sci.*, vol. 243-A, 1945.
142. Grubenmann, U., and P. Niggli, *Die Gesteinsmetamorphose*, 1924.
143. Grubenmann, U., and P. Niggli, *Metamorfizm gornykh porod (The Metamorphism of Rocks): Obshchaya chast'*, Georazvedizdat, 1933.
144. Gubkin, S. I., "Osnovnyye zakony plasticheskoy deformatsii metallov (General Laws of the Plastic Deformation of Metals)," *Stal'*, no. 11–12, 1946.
145. Gubkin, S. I., *Teoriya obrabotki metallov davleniyem (Theory of Pressure Working of Metals)*, Metallurgizdat, 1947.
146. Gutenberg, B., "Changes in Sea Level, Postglacial Uplift and Mobility of the Earth's Interior," *G. S. A. Bull.*, vol. 52, no. 5, 1941.
147. Gutenberg, B. (ed.), *Internal Constitution of the Earth*, 2d ed., New York, 1951.
148. Gutenberg, B., "The Structure of the Earth as Viewed in 1957," *Scientia*, vol. 1, 1958.
149. Gutenberg, B. (ed.), *Vnutrenneye stroyeniye zemli (The Internal Structure of the Earth)*, Izd. Inostr. Lit., 1949.
150. Gutenberg, B., "Zur Frage der Gebirgswurzeln," *Geol. Rundschau*, bd. 46, no. 1, 1957.
151. Gutenberg, B., and K. Rikhter, *Seysmichnost' zemli (The Seismic Activity of the Earth)*, Izd. Inostr. Lit., 1948.
152. Gzovskiy, M. V., "Geologicheskiy ocherk vengrii: Vengriya (A Geologic Outline of Hungary: in Hungary)," *Bol'shaya Sovietskaya Entsikl.*, 2d ed., vol. 7, 1951.
153. Gzovskiy, M. V., "Metod modelirovaniya v tektonofizike (The Method of Reproducing Tectonic Processes on Laboratory Models)," *Sov. Geof.*, no. 4, 1958.
154. Gzovskiy, M. V., "Modelirovaniye tektonicheskikh poley napryazheniy

- i razryvov (Reproduction on Models of Tectonic Fields of Stresses and Ruptures)," *Izv. A. N. S.S.S.R., ser. Geof.*, no. 6, 1954.
155. Gzovskiy, M. V., and Ye. I. Chertkova, "Modelirovaniye volnistosti prostiraniya krupnykh tektonicheskikh razryvov (Reproduction on Models of the Undulations in the Strikes of Large Tectonic Faults)," *Izv. A. N. S.S.S.R., ser. Geof.*, no. 6, 1953.
 156. Gzovskiy, M. V., "Nekotoryye osobennosti razvitiya kolebatel'nykh dvizheniy v geosinklinalyakh (Some Features of the Development of Oscillatory Movements in Geosynclines)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 6, 1948a.
 157. Gzovskiy, M. V., "O sovremennom sostoyanii nekotorykh osnovnykh ponyatiy v uchenii o geosinklinalyakh (The Present State of Certain Concepts in the Theory of Geosynclines)," *Byull. Mosk. Ob-va Isp. Pr.*, vol. 23, no. 4, 1948b.
 158. Gzovskiy, M. V., "Sootnosheniye mezhdu tektonicheskimi razryvami i napryazheniyami v zemnoy kore (Relationship between Tectonic Faults and Stresses in the Earth's Crust)," *Razvedka i Okhrana Nedr.*, no. 2, 1956.
 159. Gzovskiy, M. V., "Tektonicheskiye polya napryazheniy (Tectonic Stress Fields)," *Izv. A. N. S.S.S.R., ser. Geof.*, no. 5, 1954.
 160. Gzovskiy, M. V., "Tektonofizicheskoe obosnovaniye geologicheskikh kriteriyev seysmichnosti (Tectonophysical Basis for Geologic Signs of Seismic Activity)," *Izv. A. N. S.S.S.R., ser. Geof.*, nos. 2 and 3, 1957.
 161. Gzovskiy, M. V., "Volnistost' prostiraniya krupnykh tektonicheskikh razryvov (Undulations in the Strikes of the Planes of Large Tectonic Faults)," *Izv. A. N. S.S.S.R., ser. Geof.*, no. 2, 1953.
 162. Gzovskiy, M. V., V. N. Krestnikov, I. L. Nersesov, and G. I. Reysner, "Sopostavleniye tektoniki s seysmichnost'yu garmskogo rayona tadzhikskoy S.S.R. (A Correlation between the Tectonic Structure and Seismic Activity of the Garmskiy District of the Tadzhik S.S.R.)," *Izv. A. N. S.S.S.R., ser. Geof.*, nos. 8 and 12, 1958.
 163. Haarmann, E., *Die Oszillationstheorie*, 1930.
 164. Hall, J., *Geological Survey of New York: Vol. III. Paleontology*, 1859.
 165. Haller, J., "Probleme der tiefen tektonik: Bauformen im Migmatits sockwerke der ostgroelendischen Kaledoniden," *Geol. Rundschau*, bd. 42, h. 2, 1956.
 166. Hanson, D., "The Creep of Metals," *Trans. Am. Inst. Mining Met. Eng.*, vol. 133, 1939.
 167. Haug, E., *Geologiya (Geology)*, ONTI, 1932, vol. 1.
 168. Haug, E., "Les Géosynclinaux et les aires continentales," *Bull. Soc. géol. fr.*, sér. II, t. 28, 1900.
 169. Hedberg, H. D., "The Effect of Gravitational Compaction on the Structure of Sedimentary Rocks," *Bull. A.A.P.G.*, vol. 10, no. 2, 1926.
 170. Heezen, B. C., M. Ewing, and D. B. Ericson, "Submarine Topography in the North Atlantic," *G. S. A. Bull.*, vol. 62, no. 12, 1951.
 171. Heim, A., *Géologie der Schweiz*, bd. II, h. 1, 1919.
 172. Heim, A., *Untersuchungen über den Mechanismus der Gebirgsbildung*, bd. II, 1878.

173. Heiskanen, W. A., and F. A. Vening-Meinezs, *The Earth and Its Gravity Field*, New York, 1958.
174. Hess, H. H., "Major Structural Features of the Western North Pacific. An Interpretation of H. O. 5485, Bathymetric Chart, Korea to New Guinea," *G. S. A. Bull.*, vol. 59, no. 5, 1948.
- 174a. Hess, H. H., "Drowned Ancient Islands of the Pacific Basin," *Am. Jour. Sci.*, vol. 244, pp. 772-791, 1946.
175. Holmes, A., "Radioactivity and Earth Movements," *Trans. Geol. Soc. Glasgow*, vol. 18, pt. III, 1928-1931.
176. Holmes, A., "Some Problems of Physical Geology and the Earth's Thermal History," *Geol. Mag.*, vol. 64, 1927.
177. Hooke, R., "Lectures and Discourses on Earthquakes," in *The Posthumous Works*, London, 1705.
178. Hsieh, C., "An Outline of the Geological Structure of the Western Hills of Peiping," *Bull. Geol. Soc. China*, vol. 15, no. 1, 1936a.
179. Hsieh, C., "On the Late Mesozoic-Early Tertiary Orogenesis and Volcanism and Their Relation to the Formation of Metallic Deposits in China," *Bull. Geol. Soc. China*, vol. 15, no. 1, 1936b.
180. Hume, W., "Recent Researches on the Tertiary and Mesozoic Formations in Egypt and Sinai," *C. R. Congrès géol. int., XIII sess.*, fasc. II, 1935.
181. Ivanov, G. I., "Klivazh (otdel'nosti) v uglyakh i vmeshchayushchikh porodakh i puti yego prakticheskogo ispol'zovaniya [Cleavage (Jointing) in Coals and Their Surrounding Rocks, and Its Utilization in Practical Work]," *Tr. TSNIGRI*, t. 1, vyp. 110, 1939.
182. Ivanov, N. I., *Elementarnyy uchebnyy soprotyazheniya materialov (Elementary Textbook on the Strength of Materials)*, Gostekhizdat, 1948.
183. Joliffe, A., "Structures in the Canadian Shield," *Am. Geoph. Union Trans.*, pt. II, 1942.
184. Joly, J., *The Surface History of the Earth*, Oxford, 1925.
185. Joly, J., *Istoriya poverkhnosti zemli (The Surface History of the Earth)*, GIZ, 1929.
186. Jones, O. T., "On the Evolution of a Geosyncline," *Quart. J. Geol. Soc.*, vol. 94, pt. 2, 1938.
187. Karpinskiy, A. P., "K tektonike yevropeyskoy rossii (The Tectonics of European Russia)," *Izv. A. N.*, 1919.
188. Karpinskiy, A. P., "Obshchiy kharakter kolebaniy zemnoy kory v pre-delakh yevropeyskoy rossii (The General Nature of the Oscillatory Crustal Movements in European Russia)," *Izv. A. N.*, no. 1, 1894.
189. Karpinskiy, A. P., "Ocherk fiziko-geograficheskikh usloviy yevropeyskoy rossii v minuvskiye geologicheskiye periody (An Outline of the Physical Geography of European Russia in Past Geologic Periods)," *Zap. A. N.*, vol. 60, pril. 8, 1887.
190. Karpinskiy, A. P., "Zamechaniya o kharaktere dislokatsiy porod v yuzhnoy polovine yevropeyskoy rossii (Notes on the Nature of the Dislocations in Rocks of the Southern Half of European Russia)," *Gorn. Zhurn.*, vol. III, 1883.
191. Kassin, N. G., "Svyaz' vulkanizma i metallogeneza s tektonicheskimi strukturami kazakhstana (Volcanic Activity and Metallogenesis as Re-

- lated to Tectonic Structure in Kazakhstan)," *Probl. Sov. Geol.*, no. 8, 1937.
192. Kazakov, A. V., "Geotektonika i formirovaniye fosforitnykh mestorozhdeniy (The Geotectonics and the Formation of Phosphate Rock Deposits)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 5, 1950.
 193. Keller, B. M., "Flishevaya formatsiya paleozoya v zilayrskom sinklinorii na yuzhnom urale i skhodnyye s ney obrazovaniya (The Paleozoic Flysch in the Zilayr Synclinorium of the Southern Urals and Similar Formations)," *Tr. In-ta Geol. Nauk A. N. S.S.S.R.*, vyp. 104, 1949.
 194. Keller, B. M., "O znachenii moshchnostey pri tektonicheskikh postroyeniyakh (The Significance of the Thicknesses of Deposits in Tectonic Studies)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 6, 1948.
 195. Keller, B. M., "Verkhnelovyye otlozheniya zapadnogo kavkaza (The Upper Cretaceous Deposits of the Western Caucasus)," *Tr. In-ta Geol. Nauk A. N. S.S.S.R.*, ser. Geol., vyp. 48, no. 15, 1947.
 196. Keller, B. M., and V. V. Menner, "Paleogenovyye opolzni i svyazannyye s nimi podvodnyye opolzni (Paleogene Deposits and the Submarine Slides Associated with Them)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 20, no. 1-2, 1945.
 197. Keller, G., "Die Frage der paläogeographischen Voraussetzungen für den tektonischen Bau des Sauerlandes im Vergleich zum Ruhrgebiet," *Geol. Rundschau*, bd. 31, h. 5/6, 1940.
 198. Khabakov, A. V., *Ob osnovnykh voprosakh istorii razvitiya poverkhnosti luny (On the Fundamental Problems in the Development of the Moon's Surface)*, Izd. Mosk. Ob-va Isp. Prirody, 1949.
 199. Khabakov, A. V., *Ob osnovnykh voprosakh istorii razvitiya poverkhnosti luny (Fundamental Problems of the History of the Moon's Surface)*, Izd. Mosk. Ob-va Isp. Prirody, 1950.
 200. Khabakov, A. V., *Ocherki po istorii geologo-razvedochnykh znaniy v rossii (Notes on the History of Geological Exploration in Russia)*, Izd. MOIP, 1950, Part I.
 201. Khain, V. Ye., *Geotektonicheskiye osnovy poiskov nefi (The Geotectonic Foundations of Oil Prospecting)*, 1954.
 202. Khain, V. Ye., *Geotektonicheskoye razvitiye yugo-vostochnogo kavkaza (The Geotectonic Development of the Southeastern Caucasus)*, Aznefteizdat, 1950a.
 203. Khain, V. Ye., "Glavneyshiye cherty tektonicheskogo stroyeniya kavkaza (Main Features of the Tectonic Structure of the Caucasus)," *Sov. Geol.*, vol. 39, 1949.
 204. Khain, V. Ye., "K klassifikatsii podnyatiy i progibov geosinklinal'nykh oblastey v zavisimosti ot istorii ikh razvitiya (Classification of Uplifts and Depressions in Geosynclinal Regions in Relation to the History of Their Development)," *DAN Az.S.S.R.*, vol. 5, no. 8, 1949.
 205. Khain, V. Ye., "K probleme sootnosheniy orogeneza i epeyrogeneza (On the Relationship between Orogeny and Epeirogeny)," *Sov. Geol.*, no. 7, 1938.
 206. Khain, V. Ye., "O glybova-volnovoy (skladchato-glybovoy) strukture

- zemnoy kory [The Block and Wave (Folded-block) Structure of the Earth's Crust]," *Mosk. Ob-va. Ispyt. Prirody, Otd. Geol.*, vol. 33, no. 4, 1958.
207. Khain, V. Ye., "O nekotorykh osnovnykh ponyatiyakh v uchenii o fatsiyakh i formatsiyakh (On Some Fundamental Concepts in the Theory of Facies and Formations)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 25, no. 6, 1950b.
208. Khain, V. Ye., "O nepreryvno-preryvistom techenii tektonicheskikh protsessov (On the Continuous and Intermittent Course of Tectonic Processes)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 6, 1950c.
209. Khain, V. Ye., "Osnovnyye zakonomernosti razvitiya geosinklinaly (The Basic Laws Governing the Development of Geosynclines)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 6, 1946.
210. Khain, V. Ye., "Ostsillyatsionnyy ritm zemnoy kory (The Oscillatory Rhythm of the Earth's Crust)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 17, no. 1, 1939.
211. Khain, V. Ye., and L. N. Leont'yev, "Osnovnyye etapy geotektonicheskogo razvitiya kavkaza (Main Stages in the Geotectonic History of the Caucasus): I-II," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 25, nos. 3 and 4, 1950.
212. Khain, V. Ye., and A. N. Shardanov, "Geologicheskaya istoriya kurinskoy vpadiny v kaynozoye (The Geologic History of the Kura Depression in the Cenozoic Era)," *Izv. A. N. Az.S.S.R.*, no. 5, 1950.
213. Khlopin, V. G., "Radioaktivnost' i teplovoy rezhim zemli (The Radioactivity and the Thermal Regime of the Earth)," *Izv. A. N. S.S.S.R., Otd. Matem. i Yest. Nauk*, 1937.
214. Khuan Tsi-tsyn, *Osnovnyye cherty tektonicheskogo stroyeniya kitaya (Main Features of the Tectonic Structure of China)*, Moscow, 1952.
215. King, P., *An Outline of the Structural Geology of the United States: Guide for the International Geological Congress*, Washington, 1933.
216. Kirillova, I. V., "Nekotoryye voprosy mekhanizma skladkoobrazovaniya (Some Problems of the Mechanics of Folding)," *Tr. GEOFIAN*, vol. 6, 1949.
217. Kizeval'ter, D. S., "O stroyenii i razvitii peredovogo khrebt severnogo kavkaza (The Structure and Development of the Northern Caucasus Front Range)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 6, 1948.
218. Klenova, M. V., *Geologiya morya (Submarine Geology)*, Uchpedgiz, 1948.
219. Knopf, E. B., and E. Ingerson, *Structural Petrology*, G. S. A. Memoir 6, 1938.
220. Kobayashi, T., "The Sakana Orogenic Cycle and Its Bearing on the Origin of the Japanese Islands," *J. Faculty Sci. Imp. Univ. Tokyo*, sec. II, vol. 5, pt. 7, 1941.
221. Kober, L., *Der Bau der Erde*, 1926.
222. Kosminskaya, I. P., "Stroyeniye zemnoy kory po seysmicheskim dannym (Structure of the Earth's Crust as Determined from Seismic Data)," *Byull. Mosk. Ob-va Ispyt. Prirody, Otd. Geol.*, vol. 33, no. 4, 1958.

223. Kossmat, F., "Die mediterranean Kettengebirge in ihrer Berichtigung zum Gleichgewichtszustande der Erdrinde," *Abh. d. Sächs. Akad. d. Wiss., Math.-phys. Kl.*, no. 2, 1921.
224. Kossmat, F., Gliederung des variscischen Gebirgsbaus, *Abh. d. Sächs. Geol. Landesanstalt*, 1927.
225. Kossmat, F., *Palaeogeographie und Tektonik*, 1936.
226. Kosygin, Yu. A., "Solyanaya i gipsovaya tektonika aktyubinskoy oblasti (The Salt and Gypsum Tectonics of the Aktyubinsk Region)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 4, 1940.
227. Krashennikov, G. F., "Moshchnyye plasty uglya i differentsial'nyye tektonicheskiye dvizheniya (Thick Coal Seams and Differential Tectonic Movements)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 3, 1945.
228. Krashennikov, G. F., *Usloviya nakopleniya ugleunosnykh formatsiy S.S.S.R. (The Conditions of the Accumulation of the Coal Measures in the U.S.S.R.)*, MGU, 1957.
229. Kraus, E., *Verbiegende Baugeschichte der Gebirge*, Berlin, 1951.
230. Krenkel, E., *Geologie Afrikas*, vols. I-III, 1925-1934.
231. Krestnikov, V. N., "Istoriya razvitiya struktury i seysmichnost' tyan'shanya (History of the Structure and Seismic Activity of the Tien Shan)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 3, 1954.
232. Kropotkin, P. N., "Stroyeniye skladchatogo fundamenta tsentral'nogo kazakhstana (The Structure of the Folded Basement of Central Kazakhstan)," *Tr. In-ta Geol. Nauk A. N. S.S.S.R., ser. Geol.*, vyp. 108, no. 36, 1950.
233. Kropotkin, P. N., "O proiskhozhdenii skladchatosti (On the Origin of Folding)," *Byull. Mosk. Ob-va Isp. Pr., ser. Geol.*, vol. 25, no. 5, 1950.
234. Kropotkin, P. N., "Kosmogonicheskaya teoriya O. Yu. Shmidta i stroyeniye zemli (The Cosmogenic Theory of O. Yu. Schmidt and the Structure of the Earth)," *Izv. A. N. S.S.S.R., ser. Geogr. i Geof.*, no. 1, 1950.
235. Kropotkin, P. N., "Znachenkiye tektonicheskikh protsessov dlya obrazovaniya kislykh magm (The Effect of Tectonic Processes on the Formation of Acidic Magma)," *Tr. In-ta Geol. Nauk A. N. S.S.S.R.*, vyp. 47, 1941.
236. Kropotkin, P. N., Ye. N. Lyustikh, and N. N. Povolo-Shveykovskaya, *Anomalii sily tyazhesti na materikakh i okeanakh i ikh znachenie dlya Geotektoniki (Gravitational Anomalies over the Continents and Oceans and Their Significance in Geotectonics)*, Mosk. Gos. Universitet, Moscow, 1958.
237. Kuenen, P. H., *Submarine Geology*, New York, 1950.
238. Kuhn, W., and A. Rittmann, "Über den Zustand des Erdinnern und seine entstehung aus einem homogenen Urzustande," *Geol. Rundschau*, vol. 32, 1941.
239. Kupletskiy, B. M., "Obzor sovremennykh vzglyadov na proiskhozhdeniye granitov (A Survey of the Current Views on the Origin of Granites)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 3, 1942.
240. Kupletskiy, B. M. (ed.), *Problema obrazovaniya granitov (The Problem of Granite Formation)*, sb. 1, Izd. Inostr. Lit., 1949.
241. Kuznetsov, I. G., "Kolebatel'nye dvizheniya zemnoy kory i ikh rol' v

- strukture kavkaza (Oscillatory Crustal Movements as a Factor in the Structure of the Caucasus)," *Probl. Sov. Geol.*, no. 7, 1933.
242. Kuznetsov, I. G., "Tektonika, vulkanizm i etapy formirovaniya struktury tsentral'nogo kavkaza (Tectonics, Volcanic Activity and the Stages in the Formation of the Structure of the Central Caucasus)," *Tr. In-ta Geol. Nauk A. N. S.S.S.R.*, vyp. 131, 1951.
243. Kuznetsov, V. A., "Tektonika zapadnoy tuvy na styke s gornym altayem (The Tectonics of Western Tuva Where It Abuts on the Gornyi Altay)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 1, 1948.
244. Ladd, H. S., E. Ingerson, R. C. Townsend, M. Russell, and H. K. Stephenson, "Drilling on Eniwetok Atoll, Marshall Islands," *Bull. A.A.P.G.*, vol. 37, no. 10, 1953.
245. Lamakin, V. V., "Ob izuchenii chetvertichnykh dvizheniy zemnoy kory v oblasti pechorskoy ravniny (Quaternary Movements of the Earth's Crust in the Plain of the Pechora River)," *DAN*, vol. 62, no. 5, 1948.
246. Lamakin, V. V., "Sovremennoye podnyatiye zemnoy poverkhnosti na sredney pechore (The Current Uplift of the Earth's Surface in the Middle Pechora Region)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 4, 1945.
247. Lamplugh, G. W., "The Structure of the Weald and Analogous Tracts," *Quart. J. Geol. Soc.*, vol. 75, 1919.
248. Lawson, A. C., *The California Earthquake of April 18, 1906*, Washington, 1908.
249. Lebedeva, N. B., "Modelirovaniye protsessa obrazovaniya diapirovykh kupol (Reproduction on Models of the Formation of Diapir Domes)," *Sov. Geol.*, sb. 54, 1956.
250. Lebedeva, N. B., "O mekhanizme obrazovaniya glinyanykh diapirov (The Mechanics of the Formation of Clay Diapirs)," *Sov. Geol.*, no. 2, 1958.
251. Lee, J. S., *Geology of China*, 1939.
252. Leith, C. K., *Strukturnaya geologiya (Structural Geology)*, ONTI, 1935.
253. Leonardo da Vinci, *Notebooks*, McCurdy (tr.), London, 1938.
254. Leuchs, K., "Geologie von Asien: bd. I, t. 2. Zentralasien," in *Geologie der Erde*, 1937.
255. Levin, B. Yu., "Nekotoryye voprosy razvitiya, stroyeniya i sostava zemli (Some Problems of the History, Structure and Composition of the Earth)," *Izv. A. N. S.S.S.R., ser. Geogr.*, no. 4, 1953.
256. Levinson-Lessing, F. Yu., "K voprosu o genezise izverzhennykh porod (On the Genesis of Eruptive Rocks)," *Tr. Miner. Muzeya A. N. S.S.S.R.*, vol. 3, 1928.
257. Levinson-Lessing, F. Yu., "Problemy magmy (Problems of Magma): Part I," *Uch. Zap. Len. Gos. Univ.*, 1938.
258. Levinson-Lessing, F. Yu., "Problemy magmy (Problems of Magma): Part II," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 1, 1939.
259. Lindberg, G. U., *Chetvertichnyy period v svete biogeograficheskikh dannykh (The Quaternary Period in the Light of Biogeographical data)*, A. N. S.S.S.R., Moscow, 1955.
260. Lindberg, G. U., "Geomorfologiya okrainnykh morey vostochnoy azii i rasprostraneniye presnovodnykh ryb (The Geomorphology of the Seas

- Surrounding Eastern Asia and the Distribution of Fresh-water Fish)," *Izv. Vses. Geogr. Ob-va*, vol. 78, no. 3, 1946.
261. Lindberg, G. U., "Sovremennoye sostoyaniye problemy proiskhozhdeniya podvodnykh dolin (The Present Status of the Problem of the Origins of Submarine Canyons)," *Voprosy Geografii*, vol. 3, 1947.
 262. Li Sy-guan, *Geologiya kitaya (The Geology of China)*, Moscow, 1952.
 263. Little, O. H., "Note on the Neogene Formations of Egypt Along the Northern Part of the Red Sea," *C. R. Congrès géol. int., XIII sess.*, fasc. II, 1925.
 264. Lodochnikov, V. N., "Nekotoryye obshchiye voprosy, svyazannyye s magmoy, dayushchey bazal'tovyye porody (Some General Problems Associated with Basaltic Magma)," *Zap. Miner. Ob-va*, ser. II, ch. 68, vyp. 2 and 3, 1939.
 265. Lomonosov, M. V., *O sloyakh zemnykh (On Strata of the Earth)*, Gosgeolizdat, 1949.
 266. Lyell, C., *Osnovnyye nachala geologii (Principles of Geology)*, translated from the 9th English edition, 1866.
 267. Lyubimova, Ye. A., "Termicheskaya istoriya i temperatura zemli (The Thermal History and Temperature of the Earth)," *Byull. Mosk. Ob-va Ispytat. Prirody, Otd. Geol.*, vol. 33, no. 4, 1958.
 268. Lyustikh, Ye. N., "O vozmozhnosti ispol'zovaniya teorii akad. O. Yu. Shmidta v geotektonike (On the Possibility of Using Acad. O. Yu. Schmidt's Theory in Geotectonics)," *DAN*, vol. 59, no. 8, 1948a.
 269. Lyustikh, Ye. N., "Gravimetricheskii metod izucheniya prichin kolebatel'nykh dvizheniy zemnoy kory i nekotoryye rezul'taty yego primeneniya (The Gravimetric Method of Studying Oscillatory Crustal Movements and Some Results of Its Application)," *Izv. A. N. S.S.S.R.*, ser. Geol., no. 6, 1948b.
 270. Lyustikh, Ye. N., "Usloviya podobiya pri modelirovanii tektonicheskikh protsessov (Conditions of Similarity in Modeling Tectonic Processes)," *DAN*, vol. 64, no. 5, 1949.
 271. Lyustikh, Ye. N., "K voprosu o mekhanizme skladkoobrazovaniya (The Mechanics of Folding)," *DAN*, vol. 65, no. 6, 1949.
 272. Magnitskiy, V. A., "K voprosu o plotnosti i szhimayemosti obolochki zemli (The Density and Compressibility of the Earth's Mantle)," *Voprosy Kosmogonii*, vol. 1, 1952.
 273. Magnitskiy, V. A., "O vozmozhnom kharaktere deformatsiy v glubokikh sloyakh zemnoy kory i podkorovom sloye (The Possible Nature of the Deformations in the Depths of the Earth's Crust and the Subcrustal Layer)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 23, no. 2, 1948.
 274. Magnitskiy, V. A., *Osnovy fiziki zemli (Fundamentals of Earth Physics)*, Moscow, 1953.
 275. Makovelskiy, A., *Dosokratiki: Ch. 1 (The Presocratics: Part 1)*, 1914.
 276. Makowski, A., "Über die faunistischen Horizonte und die Oscillationerscheinungen im Rybniker Karbon," *C. R. II Congrès pour l'avancement des études de stratigraphie carbonifère (Heerlen, 1935)*, t. II, 1937.

277. *Materialy po geologii tsentral'nogo kazakhstana (Materials on the Geology of Central Kazakhstan)*, A. N. S.S.S.R., 1940.
278. Mazarovich, A. N., "O plashcheobraznom zaleganii v oblasti povolzh'ya (Stratified Deposits in the Volga Region)," *Geol. Vestnik*, no. 4, 1921.
279. McDonald, D. F. M., "Some Factors of Central American Geology," *Bull. A. A. P. G.*, vol. 4, no. 3, 1920.
280. Menard, H. W., "Pleistocene and Recent Sediments from the Floor of the Northeastern Pacific Ocean," *G. S. A. Bull.*, vol. 64, no. 11, 1953.
281. Menard, H. W., "Shear Zones of the Northeastern Pacific Ocean and Anomalous Structural Trends of Western North America," *G. S. A. Bull.*, vol. 64, no. 12, pt. 2, 1953 (abstr.).
- 281a. Menard, H. W., "Deformation of the Northeastern Pacific Basin and West Coast of North America," *Bull. Geol. Soc. Am.*, vol. 66, pp. 1149-1198, 1955.
282. Menard, H. W., and R. S. Dietz, "The Mendocino Submarine Escarpment," *J. Geol.*, vol. 60, no. 3, 1952.
283. Meshcheryakov, Yu. A., "Ob osnovnykh zakonmernostyakh stroyeniya i razvitiya krupnykh form rel'yefa russkoy ravniny (The Basic Regular Features in the Structure and Development of the Major Elements in the Relief of the Russian Plain)," *DAN*, vol. 79, no. 1, 1951a.
284. Meshcheryakov, Yu. A., "Ob otrazhenii v rel'yefe russkoy ravniny antiklinal'nykh struktur tipa valov i kupolov (The Reflection of Anticlinal Structures Such as Swells and Domes in the Relief of the Russian Plain)," *DAN*, vol. 79, no. 2, 1951b.
285. Meshcheryakov, Yu. A., "O morfologicheskoy strukture severo-zapada russkoy ravniny (On the Morphological Structure of the Northwestern Part of the Russian Plain)," *Izv. A. N. S.S.S.R., ser. Geogr. i Geof.*, vol. 14, no. 5, 1950.
286. Meshcheryakov, Yu. A., "Sovremennyye dvizheniya zemnoy kory (Present Movements of the Earth's Crust)," *Priroda*, no. 9, 1958.
287. Meshcheryakov, Yu. A., and M. I. Sinyagina, "Opyt izucheniya sovremennykh dvizheniy zemnoy kory po daniym povtornogo nivelirovaniya (An Attempt to Study the Current Crustal Movements by Repeated Leveling)," *Izv. A. N. S.S.S.R., ser. Geogr.*, no. 1, 1951.
288. Mikhaylov, A. E., "Osnovnyye etapy razvitiya predkarpatskogo krayevogo progiba (The Main Stages in the Development of the Ciscarpathian Foredeep)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 26, no. 3, 1951.
289. Milanovskiy, Ye. V., "Ocherk teorii geosinklinal'ey v yeye sovremennom sostoyanii (An Outline of the Theory of Geosynclines in Its Present Status)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 7, no. 4, 1929.
290. Milanovskiy, Ye. V., "Ponyatiya orogeneza i epirogeneza v kriticheskom osveshchenii Stille (The Concepts of Orogenesis and Epeirogenesis in a Critical Evaluation of Stille's Views)," *Vestn. MGA*, vol. 1, 1923.
291. Milanovskiy, Ye. V., "Teoriya orogenicheskikh protsessov kviringa (Kviring's Theory of Orogenic Processes)," *Vestn. MGA*, vol. 2, 1924.
292. Misch, P., "Der Bau der mittleren Südpynaeen," *Abh. ges. Wiss. zu Goettingen, Math-phys. Kl., folge III*, h. 12, 1936.

293. Molchanov, I. A., "Geometricheskiy analiz postupatel'nykh diz'yunktivov (Geometrical Analysis of Advancing Disjunctive Dislocations)," in *Marksheyderskiy Sbornik*, vol. II, Tomsk, 1935.
294. Moro, A. L., *Dei crustacei*, Venice, 1740.
295. Mrazec, L., "Les Plis diapires," *C. R. Séances Inst. géol. Roumanie*, vol. 2, 1915.
296. Muratov, M. V., "Tektonika i istoriya razvitiya al'piyskoy geosinklinal'noy oblasti yuga yevropeyskoy chasti S.S.S.R. i sopredel'nykh stran (The Tectonics and the History of the Geosynclinal Region in the South of the European Part of the U.S.S.R. and the Adjacent Countries)," in *Tektonika S.S.S.R.*, vol. 2, A. N. S.S.S.R., 1949.
297. Muratov, M. V., "Yurskiye porfirityvyye intruzii v verkhov'yakh rek khasauta i eshkakona (severnyy kavkaz) (Jurassic Porphyrite Intrusives on the Upper Reaches of the Khasaut and Eshkakon Rivers in the Northern Caucasus)," *Izv. A. N. S.S.S.R.*, ser. Geol., no. 3, 1945.
298. Murray, H. W., "Profiles of the Aleutian Trench," *G. S. A. Bull.*, vol. 56, no. 7, 1945.
299. Murray, H. W., "Submarine Mountains in the Gulf of Alaska," *G. S. A. Bull.*, vol. 52, no. 3, 1941.
300. Mushketov, I. V., *Fizicheskaya geologiya (Physical Geology)*, vol. 1, 4th ed., ONTI, 1935.
301. Mushketov, I. V., "Vernenskoye zemletryaseniye 1887 g. 28 maya (The Verny Earthquake of May 28, 1887)," *Tr. Geol. Kom.*, vol. 10, no. 1, 1890.
302. Nalivkin, D. V., "Skladchatost' i nesoglasiye (Folding and Unconformity)," in *Sbornik Akademiku V. I. Vernadskomu*, A. N. S.S.S.R., 1936.
303. Nettleton, L. L., "Fluid Mechanics of Salt Domes," *Bull. A. A. P. G.*, vol. 18, no. 9, 1934.
304. Nettleton, L. L., "Recent Experimental and Geophysical Evidence of the Mechanics of Salt Dome Formation," *Bull. A. A. P. G.*, vol. 27, no. 1, 1943.
305. Nevin, C. M., *Principles of Structural Geology*, 3d ed., 1942.
306. Niggli, P., *Mineral- und Gesteinsprovinzen*, 1923.
307. Nikolayev, N. I., "Noveyshaya tektonika S.S.S.R. (The Recent Tectonics of the U.S.S.R.)," *Tr. Komiss. Po. Izuch. Chetvertichn. Perioda*, A. N. S.S.S.R., vol. 8, 1949.
308. Noelke, F., *Geotektonische Hypothesen*, 1924.
309. Novikova, A. C., "O treshchinovatosti osadochnykh porod vostochnoy chasti russkoy platformy (Jointing in the Sedimentary Rocks of the Eastern Part of the Russian Platform)," *Izv. A. N. S.S.S.R.*, ser. Geol., no. 5, 1951.
310. O'Brien, C. A. E., "Salztektonik in Südpersien," *Zeitschr. Deutsch. geol. Ges.*, bd. 105, no. 4, 1953.
311. O'Brien, C. A. E., "Tectonic Problems of the Oilfield Belt of Southwest Iran," *Rep. XVII Sess. Int. Geol. Congress*, 1948, pt. VI, London, 1950.
312. Obruchev, V. A., *Geologicheskii obzor sibiri (An Outline of the Geology of Siberia)*, GIZ, 1927.

313. Obruchev, V. A., *Geologiya sibir* (*The Geology of Siberia*), vols. I–III, Izd. A. N. S.S.S.R., 1935–1938.
314. Obruchev, V. A., "Novye techeniya v geotektonike (New Currents in Geotectonics)," *Izv. Geol. Kom.*, no. 3, 1926.
315. Obruchev, V. A., "Pul'satsionnaya gipoteza geotektoniki (The Pulsation Hypothesis in Geotectonics)," *Izv. A. N. S.S.S.R., ser. Geol., vyp. 1*, 1940.
316. Officer, C. B., M. Ewing, and P. C. Wuenschell, "Seismic Refraction Measurements in the Atlantic Ocean: Part IV," *G. S. A. Bull.*, vol. 63, no. 8, 1952.
317. Oliver, J. E., and C. L. Drake, "Geophysical Investigations in the Emerged and Submerged Atlantic Coastal Plain: Part VI," *G. S. A. Bull.*, vol. 62, no. 11, 1952.
318. Panov, D. G., "K probleme podvodnykh kan'onov (The Problem of Submarine Canyons)," *Zemlevedeniye*, vol. 2, no. 42, 1948.
319. Panov, D. G., "O proiskhozhdenii i istorii razvitiya okeanov (The Origin and the History of the Oceans)," *Voprosy Geografii*, vol. 12, 1949.
320. Panov, D. G., "Problema proiskhozhdeniya materikov i okeanov v svete novykh issledovaniy (The Origin of the Continents and the Oceans in the Light of Recent Investigations)," *Priroda*, no. 3, 1950.
321. Pariyskiy, N. N., "Nepostoyanstvo vrashcheniya zemli i eye deformatsiya (The Nonuniformity of the Earth's Rotation and Its Deformation)," in *Tr. Soveshch. po Metodam Izucheniya Dvizheniy i Deformatsiy Zemnoy Kory*, Geodezizdat, 1948.
322. Pavlinov, V. N., "K voprosu o klassifikatsii intruzivnykh tel (On the Problem of the Classification of Intrusive Bodies)," *Voprosy Teoretich. i Prikl. Geologii (MGRI)*, vol. 2, 1947.
323. Pavlinov, V. I., "Osnovnyye zakonomernosti stroyeniya i obrazovaniya malykh intruziy tipa lakkolitov (The Main Features of the Structure and Formation of Minor Intrusives of the Laccolith Type)," *Tr. MGRI*, vol. 23, 1946.
324. Pavlinov, V. N., "O strukture nekotorykh lakkolitov rayona kavkazskikh mineral'nykh vod (On the Structure of Certain Laccoliths in the Region of the Caucasus Mineral Waters)," *Byull. Mosk. Ob-va Isp. Pr.*, vol. 21, no. 2, 1946.
325. Pavlov, A. P., "Robert Hooke: Un évolutionniste oublié du XVII siècle," *Palaeobiologica*, bd. 1, 1928.
326. Pavlov, A. P., "Samarskaya luka i zhiguli (Samarskaya Luka and Zhiguli)," *Tr. Geol. Kom.*, vol. 2, no. 5, 1887.
327. Pavlovskiy, Ye. V., "Sravnitel'naya tektonika mezozoyskikh struktur vostochnoy sibir i velikogo rifta afriki i aravii (A Comparison of the Tectonics of the Mesozoic Structures in Eastern Siberia and the Great Rift Valleys of Africa and Arabia)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 5, 1948.
328. Pek, A. V., *Treshchinnaya tektonika i strukturnyy analiz (Structural Analysis and the Tectonics of Jointing)*, Izd. A. N. S.S.S.R., 1939.
329. Pek, A. V., "O Mekhanizme vozniknoveniya slantsevatosti (The Me-

- chanics of the Formation of Slaty Cleavage)," *Izv. A. N. S.S.S.R., ser. Geol.*, vyp. 2, 1940.
330. Pekeris, C. L., "Thermal Convections in the Interior of the Earth," *Monthly Notices, Roy. Astron. Soc., Geophys. Suppl.*, vol. 3, pp. 343-367, 1935.
 331. Permyakov, Ye. N., "Tektonicheskaya treshchinovatost' russkoy platformy (The Tectonic Jointing of the Russian Platform)," *Mat-ly k Poznaniyu Geol. Stroen. S.S.S.R., Izdavayemye Mosk. Obshch. Ispytat. Prirody*, vyp. 12, 1949.
 332. Peshl', T., *Soprotivleniye materialov (Strength of Materials)*, Gos. Izd. Tekhniko-teoretich. Lit-ry, 1948.
 333. Petrushevskiy, B. A., "K istorii razvitiya tyan'-shanya v mezozoysskoye i kaynozoysskoye vremya (On the History of the Tien Shan in Mesozoic and Cenozoic Times)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 23, no. 5, 1948a.
 334. Petrushevskiy, B. A., "Stroyeniye tretichnykh otlozheniy tyan'-shanya (The Structure of the Tertiary Deposits in the Tien Shan)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 23, no. 1, 1948b.
 335. Petrushevskiy, B. A., "Mezozoyssko-kaynozoysskaya struktura zapadno-sibirskoy nizmennosti (The Mesozoic and Cenozoic Structure of the Western Siberian Lowland)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 26, no. 4, 1951.
 336. Petrushevskiy, B. A., *Uralo-sibirskaya epigertsinskaya platforma i tyan'-shan' (The Ural-Siberian Epihercynian Platform and the Tien Shan)*, A. N. S.S.S.R., 1955.
 337. Petrushevskiy, B. A., "Znachenkiye geologicheskikh yavleniy pri seysmicheskoy rayonirovaniy (The Importance of Geologic Phenomena in Seismic Regionalization)," *Trudy Geof. In-ta A. N. S.S.S.R.*, no. 28, p. 155, 1955.
 338. Peyve, A. V., "Asimmetriya glubinnykh tektonicheskikh struktur uralo-tyan'shan'skogo orogena i proiskhozhdeniye ego virgatsiy (The Asymmetry of the Deep Tectonic Structures of the Ural-Tien Shan Orogenesis and the Origin of Its Virgation)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 21, no. 5, 1947a.
 339. Peyve, A. V., "Glubinnye razlomy v geosinklinal'nykh oblastiakh (Deep Faults in Geosynclinal Regions)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 5, 1945.
 340. Peyve, A. V., "Obshchaya kharakteristika, klassifikatsiya i prostranstvennoye raspolozheniye glubinnykh razlomov (The General Features, the Classification and the Distribution in Space of Deep-seated Faults)," *Izv. A. N. S.S.S.R., ser. Geol.* nos. 1 and 3, 1956.
 341. Peyve, A. V., "O 'zakone' inversii v geologii kavkaza (On the 'Law' of Inversion in the Geology of the Caucasus)," *Sov. Geol.*, vyp. 4, 1941.
 342. Peyve, A. V., "Tektonika severoural'skogo boksitovogo poyasa (The Tectonics of the Northern Urals Bauxite Zone)," *Mat-ly k Pozn. Geol. Stroen. S.S.S.R. (MOIP)*, vol. 4, no. 8, 1947b.
 343. Peyve, A. V., "Tipy i razvitiye paleozoysskikh struktur uralo-tyan'shan'-

- skey geosinklinal'noy oblasti (The Types and the Development of the Paleozoic Structures in the Ural-Tien Shan Geosynclinal Region)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 6, 1948.
344. Peyve, A. V., and V. M. Sinitsyn, "Nekotoryye osnovnyye voprosy ucheniya o geosinklinalyakh (Some Fundamental Problems in the Theory of Geosynclines)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 4, 1950.
 345. Picard, L., "Outlines of the Tectonics of the Earth," *Bull. Geol. Dep., Hebrew Univ., Jerusalem*, vol. 2, no. 3-4, 1936.
 346. Playfair, J., *Illustrations of the Huttonian Theory of the Earth*, 1802.
 347. Poldervaart, A., *The Crust of the Earth (A Symposium)*, G. S. A. Special Paper 62, New York, 1955.
 348. Polkanov, A. A., *Geologicheskiye issledovaniya v rayonakh magmaticheskikh i metamorficheskikh porod (Geologic Investigations in Regions of Igneous and Metamorphic Rocks)*, 1934.
 349. Popov, V. I., *Istoriya depressiy i podnyatiy zapadnogo tyan'shanya (The History of the Depressions and Uplifts in the Western Tien Shan)*, Tashkent, 1938.
 350. Pratt, J., "On the Attraction of the Himalaya Mountains," *Phil. Trans. Roy. Soc. (London)*, vol. 145, 1855.
 351. "Problem of Land Connections across the South Atlantic with Special Reference to the Mesozoic," *Bull. Am. Museum Nat. Hist.*, vol. 99, art. 3, New York, 1952.
 352. Pruvost, P., *Sédimentation et subsidence: Livre jubilaire centenaire de la Soc. géol. de France, 1830-1930*, Paris, 1930, vol. 2.
 353. Pustovalov, L. V., *Petrografiya osadochnykh porod (The Petrography of Sedimentary Rocks)*, Gostoptekhizdat, 1940.
 354. Ramberg, H., "Radial Diffusion and Chemical Stability in the Gravitational Field," *J. Geol.*, vol. 56, no. 5, 1948.
 355. Ramsey, W. H., On the Nature of the Earth's Core, *M. N. Geophys.*, suppl. 5, no. 9, 1949.
 356. Rebinder, P. A., "Fiziko-khimicheskiye issledovaniya protsessov deformatsii tverdykh tel (Physicochemical Investigations of the Deformation of Solid Bodies)," in *Yubileyn. Sborn., Posvyashch. 30-Letiye Velikoy Okt. Sots. Rev.*, A. N. S.S.S.R., 1947.
 357. Reed, R. D., *Geology of California*, 1926.
 358. Reed, F. R. C., *The Geology of the British Empire*, 1921.
 359. Rellum, L. B., R. W. Imlay, and U. G. Kane, "Evolution of the Coahuila Peninsula, Mexico," *G. S. A. Bull.*, vol. 47, no. 7, 1936.
 360. Rengarten, V. P., "Tektonicheskaya kharakteristika skladchatykh oblastey kavkaza (The Tectonic Features of the Folded Regions of the Caucasus)," *Tr. Tret'yego S'ezda Geologov*, vyp. 2, 1930.
 361. Reyer, E., *Teoretische Geologie*, 1888.
 362. Rezanov, I. A., "Ashkhabadskoye zemletryaseniye 1948 g. i geologicheskiye usloviya yego vozniknoveniya (The Ashkhabad Earthquake of 1948 and the Geologic Conditions Causing It)," *Izv. A. N. S.S.S.R., ser. Geof.*, no. 6, 1958.
 363. Richards, H. C., "Results of Deep-boring Operations on the Great Barrier Reef, Australia," *Proc. 6th Pacific Sci. Congr.*, vol. 11, 1940.

364. Richey, J. E., *Scotland, The Tertiary Volcanic Districts: British Regional Geology*, Edinburgh, 1948.
365. Richter, G., "Das Grenzgebiet Alpen-Pyrenaeen," *Abh. ges. Wiss. zu Goettingen, Math-phys. Kl.*, folge III, h. 19, 1939.
366. Richter, G., and R. Teichmueller, "Die Entwicklung der keltiberischen Ketten," *Abh. ges. Wiss. zu Goettingen, Math-phys. Kl.*, folge III, h. 7, 1933.
367. Richter, G., "Die Rolle der epirogener Schwelle im Faltengebirge," *Geol. Rundschau*, bd. 28, h. 5, 1937.
368. Robinson, V. N., "Cherez glavnyy khrebet i kavkazskiy gosudarstvennyy zapovednik (Through the Main Caucasus Range and the Caucasus State Preserve)," *Mezhd. Geol. Kongr., XVII sessiya, ekskursiya po kavkazy, glavnyy khrebet zapovednik*, Moscow, 1937.
369. Ronov, A. B., "Istoriya osadkonakopleniya i kolebatel'nykh dvizheniy yevropeyskoy chasti S.S.S.R. (The History of Sedimentary Deposition and Oscillatory Movements in the European Part of the U.S.S.R.)," *Tr. Geof. In-ta A. N. S.S.S.R.*, no. 3, p. 130, 1949.
370. Ronov, A. B., and V. Ye. Khain, "Istoriya osadkonakopleniya v srednem i verkhnem paleozoye v svyazi s gertsinskim etapom geotektonicheskogo razvitiya zemnoy kory (The History of Sedimentation in the Middle and Late Paleozoic, in Connection with the Hercynian Stage in the Geotectonic Development of the Earth's Crust)," *Sov. Geol.*, vol. 58, 1957.
371. Rostovtsev, D. S., "Osnovnyye voprosy teorii i praktiki v oblasti upravleniya krovley po dannym zagranitsy i S.S.S.R. i ikh znachenie s tochki zreniya obshchikh interesov narodnogo khozyaystva (Basic Problems in the Theory and Practice of Topsoil Management According to Data from the U.S.S.R. and other Countries, and Their Significance from the Standpoint of the National Economy)," in *Sbornik Mat-lov po Voprosu Upravleniya Krovley*, ONTI, 1935.
372. Rozanov, L. N., *Istoriya formirovaniya tektonicheskikh struktur bashkirii i prilezhashchikh oblastey (History of the Formation of the Tectonic Structures in Bashkiria and Adjoining Regions)*, 1957.
373. Rozanov, P. N., "Kolebatel'nyye dvizheniya i formirovaniye platformennykh struktur (Oscillatory Movements and the Formation of Platform Structures)," *Sov. Geol.*, sb. 39, 1949.
374. Rubey, W. W., "Geologic History of Sea Water," *G. S. A. Bull.*, vol. 62, no. 9, 1951.
375. Ruedemann, R., and R. Balk, *Geology of North America: Vol. I. Geologie der Erde*, 1939.
376. Rukhin, L. B., *Osnovy litologii (Foundations of Lithology)*, 1953.
377. Rukhin, L. B., *Tipy peschanykh fatsiy (Types of Sandstone Facies): I. Litologicheskii Sbornik*, Gostoptekhizdat, 1948.
378. Rumyantsev, S., "Tektonicheskiye narusheniya, nablyudayushchiesya po severo-zapadnoy okraine kuznetskogo kamennougol'nogo basseyna, i ikh ob'yasneniye (Tectonic Dislocations Observed Along the North-western Margin of the Kuznets Coal Basin, and Their Explanation)," *Gorniy Zhurn.*, nos. 10 and 11, 1928.

379. Ruzhentsev, V. Ye., "Osnovy tektoniki uralo-embenskogo rayona (Basic Features of the Tectonics of the Ural-Emba Region)," *Byull. Mosk. Obva Isp. Pr., Otd. Geol.*, vyp. 1-2, 1930.
380. Sander, B., *Einführung in die Gefügekunde der geologischen Körper*, vols. I-II, Wien, 1948-1950.
381. Sapozhnikov, D. G., "Tektonika zapadnoy chasti kazakhskoy skladchatoy strany (The Tectonics of the Western Part of the Kazakhstan Folded Region)," in *Tektonika S.S.S.R.: Vol. I. Tektonika Tsentr. Kazakhstana*, Part 1, 1948.
382. Savarenskiy, Ye. F., and D. P. Kimos, *Elementy seysmologii i seysmometrii (Elements of Seismology and Seismometry)*, Tekhteorizdat, 1949.
383. Savarenskiy, Ye. F., "O seysmichenskikh volnakh, otrazhennykh na glubine 900 km (Seismic Waves, Reflected at the Depth of 900 Km)," *Tr. Seysmol. In-ta A. N. S.S.S.R.*, no. 106, 1940.
384. Savarenskiy, Ye. F., "K voprosu o neodnorodnosti v glubinnom stroenii zemli po seysmicheskim dannym (On the Nonuniformity of the Structure in the Earth's Depths, According to Seismic Data)," *DAN*, vol. 27, no. 1, 1940.
385. Schuchert, C., *Geology of the Antillean-Caribbean Region*, 1935.
386. Schuchert, C., "Orogenic Times of the Northern Appalachians," *G. S. A. Bull.*, vol. 41, no. 4, 1930.
387. Schuchert, C., "Sites and Nature of the North American Geosynclines," *G. S. A. Bull.*, vol. 34, 1923.
388. Scupin, H., "Kontinentale und marine Senkungsbecken als Ausdruck epirogenetischer Bewegungen," *Geol. Rundschau*, bd. 25, h. 1, 1934.
389. *Secular Variations of Sea Level*, Assoc. d'Océanographie Physique Union Géod. et Géophys. Intern., Publ. Sci. 13, 1954.
390. Sel'skiy, V. A., *Solyanyye kupola i ikh svyaz' s nef't'yu (Salt Domes and Their Association with Oil)*, ONTI, 1936.
391. Semenenko, N. P., *Struktura rudnykh poley krivorozhskikh zhelezorudnykh mestorozhdeniy (The Structure of the Ore Fields in the Krivoy Rog Iron-ore Deposits): 1*. Kiev, Izd. A. N. S.S.S.R., 1946.
392. Shainin, V. E., "Conjugate Sets of en échelon Tension Fractures in the Athens Limestone of Riverston, Virginia," *G. S. A. Bull.*, vol. 61, no. 6, 1950.
393. Shand, S. Y., "Rocks of the Mid-Atlantic Ridge," *J. Geol.*, vol. 57, no. 1, 1949.
394. Shatskiy, N. S., "Gipoteza vegerena i geosinklinali (Wegener's Hypothesis and Geosynclines)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 4, 1946a.
395. Shatskiy, N. S., "Bol'shoy donbass i sistema vichita: sravnitel'naya tektonika drevnykh platform (The Great Donets Basin and the Wichita System: Comparative Tectonics of Ancient Platforms), Stat'ya 2," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 6, 1946b.
396. Shatskiy, N. S., "Osnovnyye cherty stroeniya i razvitiya vostochno-evropeyskoy platformy: sravnitel'naya tektonika drevnykh platform (The Main Features of the Structure and Development of the Eastern Euro-

- pean Platform: Comparative Tectonics of Ancient Platforms), Stat'ya 1," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 1, 1946c.
397. Shatskiy, N. S., "Nekotoryye soobrazheniya o tektonike tsentral'nogo kazakhstana (Some Remarks on the Tectonics of Central Kazakhstan)," in *Mat-ly po Geol. Tsent. Kazakhstana*, A. N. S.S.S.R., 1940.
 398. Shatskiy, N. S., "Ocherki tektoniki volgo-ural'skoy neftenosnoy oblasti i smezhnoy chasti zapadnogo sklona yuzhnogo urala (Outlines of the Tectonics of the Volga-Urals Oil-bearing Region and the Adjoining Parts of the Western Slopes of the Southern Urals)," *Mat-ly k Pozn. Geol. Stroyen. S.S.S.R.*, nov. ser., vyp. 2/6, 1945a.
 399. Shatskiy, N. S., "O sravnitel'noy tektonike severnoy ameriki i vostochnoy yevropy (The Comparative Tectonics of North America and Eastern Europe)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 4, 1945b.
 400. Shatskiy, N. S., "O dlitel'nosti skladkoobrazovaniya i o fazakh skladchatosti (The Duration of the Process of Folding and the Phases of Folding)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 1, 1951.
 401. Shatskiy, N. S., "O glubokikh dislokatsiyakh, okhvatyvayushchikh platformy i skladchatyye oblasti (povolzh'ye i kavkaz) [Deep Dislocations Encompassing Platforms and Folded Regions (of the Povolzh'ye Area and the Caucasus)]," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 5, 1948.
 402. Shatskiy, N. S., "Osnovnyye cherty tektoniki sibirskoy platformy (Main Features of the Tectonics of the Siberian Platform)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 10, vyp. 3-4, 1932.
 403. Shatskiy, N. S., "O strukturnykh svyazyakh platform so skladchatymi geosinklinal'nymi oblastyami (On the Structural Relationship between Platforms and Geosynclinal Folded Regions)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 5, 1947.
 404. Shatskiy, N. S., "O tektonike tsentral'nogo kazakhstana (The Tectonics of Central Kazakhstan)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 5-6, 1938.
 405. Shatskiy, N. S., "O tektonike vostochno-yevropeyskoy platformy (On the Tectonics of the Eastern European Platform)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 15, vyp. 1, 1937a.
 406. Shatskiy, N. S., "O neokatastrofizme (Neocatastrophism)," *Probl. Sov. Geol.*, vol. 7, no. 7, 1937b.
 407. Shatskiy, N. S., "Proiskhozhdeniye donetskogo basseyna (The Origin of the Donets Basin)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 15, vyp. 4, 1937c.
 408. Shatskiy, N. S., "Stratigrafiya i tektonika verkhnemelovykh i nizhnetretichnykh otlozheniy severnoy okrainy donetskogo kryazha (The Stratigraphy and Tectonics of the Late Cretaceous and Early Tertiary Deposits along the Northern Margin of the Donets Ridge)," *Tr. OKMA*, vyp. 5, 1924.
 409. Shepard, F. P., *Submarine Geology*, 1948.
 410. Shirokov, A. Z., "Geotektonicheskiye predposylki uglianakopleniya v zapadnoy chasti donetskogo progiba (The Geotectonic Assumptions Underlying the Study of the Coal Deposits in the Western Part of the Donets Basin)," *Sov. Geol.*, vol. 38, 1949.
 411. Shirokov, A. Z., "O moshchnosti otlozheniy donetskogo karbona (The

- Thicknesses of the Deposits in the Donets Carboniferous), *Sov. Geol.*, vol. 8, no. 12, 1938.
412. Shmidt, O. Yu., *Chetyre leksii o teorii proiskhozhdeniya zemli (Four Lectures on the Theory of the Origin of the Earth)*, 2d ed., Izd. A. N. S.S.S.R., 1950.
413. Shneyerson, B. L., "O primeneni teorii podobiya pri tektonicheskom modelirovanii (Application of the Theory of Similarity in the Reproduction of Tectonic Phenomena on Models)," *Tr. Inst. Teoretich. Geof.*, vol. 3, A. N. S.S.S.R., 1947.
414. Shreyner, L. A., *Fizicheskiye osnovy mekhaniki gornykh porod (The Physical Foundations of the Mechanics of Rocks)*, Gostoptekhizdat, 1950.
415. Shtreys, N. A., "O nekotorykh osnovnykh ponyatiyakh v uchenii o geosinklinalyakh (Some Basic Concepts of the Theory of Geosynclines)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 22, no. 5, 1947.
416. Shtreys, N. A., "Stratigrafiya i tektonika zelenokamennoy polosy srednego urala (The Stratigraphy and Tectonics of the Greenstone Belt of the Central Urals)," in *Tektonika S.S.S.R.*, vol. III, A. N. S.S.S.R., 1951.
417. Shul'ts, S. S., "Analiz Noveyshey tektoniki i rel'yef tyan'shanya (Analysis of the Recent Tectonics and the Relief of the Tien Shan)," *Zap. Vses. Geogr. Ob-va*, nov. ser., vol. 3, 1948.
418. Shumilin, S. V., "O tektonike embenskogo rayona (On the Tectonics of the Emba Region)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 11, no. 1, 1933.
419. Shurygin, A. M., "Usloviya formirovaniya struktur tsentral'noy chasti yugo-vostochnogo kavkaza (Conditions of the Formation of the Structure of the Central Southeastern Caucasus)," *Sov. Geol.*, nos. 3 and 8, 1958.
420. Sinitsyn, V. M., "Osnovnyye cherty tektoniki kitaya (Basic Features of the Tectonics of China)," in *Voprosy Geologii Azii*, A. N. S.S.S.R., 1955.
421. Sinitsyn, V. M., *Tektonicheskaya karta S.S.S.R., pod redaktsiyey N. S. Shatskogo (Tectonic Map of the U.S.S.R., edited by N. S. Shatskiy)*, 1956.
422. Sinitsyn, V. M., "Stroyeniye i razvitiye kitayskoy platformy (The Structure and Development of the Chinese Platform)," *Izv. A. N. S.S.S.R.*, ser. Geol., no. 6, 1948.
423. Slesarev, V. D., *Mekhanika gornykh porod (Rock Mechanics)*, Ugletekhizdat, 1948.
424. Slesarev, V. D., *Obrusheniye i osedaniye gornykh porod (Collapse and Subsidence of Rocks)*, ONTI, 1936.
425. Smirnov, V. I., "Rudnyye poyasa (Ore Belts)," in *Voprosy Teoreticheskoy i Prikladnoy Geologii*, vyp. 1, Moscow, 1947.
426. Sneegans, D., *La Géologie des nappes de l'ubaye-embrunais entre la durance et l'ubaye, Mém. pour servir à l'expl. Carte géol. France*, Paris, 1938.
427. Solignac, M., "Description d'une nouvelle carte géologique de la Tunisie, à l'échelle de 1/500,000," *C. R. XV Sess. Int. Geol. Congr. Vol. II. South Africa*, 1929-1930.

428. Sokolovskiy, V. V., *Teoriya plastichnosti (The Theory of Plasticity)*, Izd. A. N. S.S.S.R., 1946.
429. Sokratov, G. I., "Iz istorii russkoy geologii vtoroy poloviny XIX v (The History of Russian Geology in the Second Half of the XIX Century)," *Zap. Len. Gorn. In-ta*, vols. 15–16, 1949.
430. Sonder, R. A., "Die erdgeschichtlichen Diastrophismen," *Geol. Rundschau*, vol. 13, 1922.
431. Sorskiy, A. A., "Mekhanizm obrazovaniya melkikh strukturnykh form v metanorfitskikh tolshchakh arkheya (The Mechanics of the Formation of Small Structures in Archean Metamorphic Series)," *Tr. Geofan*, no. 18, p. 145, 1952.
432. Sorskiy, A. A., "O roli posloynnykh differentsial'nykh dvizheniy pri formirovani i skladchatykh struktur (The Role of the Differential Movements along the Beds in the Formation of Folded Structures)," *DAN*, nov. ser., vol. 22, no. 3, 1950a.
433. Sorskiy, A. A., "O mekhanizme tektonicheskogo razlinzovaniya gornykh porod (On the Mechanics of the Formation of Tectonic Lenses in Rocks)," *DAN*, nov. ser., vol. 22, no. 5, 1950b.
434. Soustov, N. I. (ed.), *Problema obrazovaniya granitov (The Problem of Granite Formation)*, Izd. Inostr. Lit., 1950, vol. 2.
435. Spurr, J. E., The Relation of Ore Deposition to Faulting, *Econ. Geol.*, vol. 2, 1916.
436. Starik, I. Ye., *Radioaktivnyye metody opredeleniya geologicheskogo vremeni (Radioactive Methods of Determining Geologic Time)*, ONTI, 1938.
437. Staub, R., *Grundsätzliches zur Anordnung und Entstehung der Kettengebirge*, Wien, 1953.
438. Stearns, H. T., "Late Geologic History of the Pacific Basin," *Am. J. Sci.*, vol. 243, pp. 614–626, 1945.
439. Stille, H., "Senkungs-, Sedimentations- und Faltungsräume," *C. R. XI Congrès géol. Int.*, 1910, fasc. 2, 1912.
440. Stille, H., Injectivfaltung und damit zusammenhängende Erscheinungen, *Geol. Rundschau*, bd. 8, 1917.
441. Stille, H., "Über Hauptformen der Orogenese und ihre Verknüpfung," *Nachr. d. K. Ges. d. Wiss. in Goettingen, Math-phys. Kl.*, 1918.
442. Stille, H., *Grundfragen der vergleichenden Tektonik*, 1924.
443. Stille, H., "The Upthrust of the Salt Masses of Germany," *Bull. A. A. P. G.*, vol. 9, no. 3, 1925.
444. Stille, H., "Bemerkungen zur perimesetischen Faltung in ihrem südpyrrenäisch-balearenischen Abteile," *Abh. Ges. Wiss. zu Goettingen, Math-phys. Kl.*, folge III, h. 10, 1934.
445. Stille, H., Geologische Untersuchungen im Westlichen Mediterrangebiet, *Geol. Rundschau*, bd. 28, h. 1/2, 1937.
446. Stille, H., *Geotektonische Probleme im atlantischen Raum*, 1937.
447. Strabon, *Erdbeschreibung in siebzehn Büchern: I–III*, 1831–1833.
448. Strakhov, N. M., "Istoriko-geologicheskiye tipy osadkonakopleniya (Types of Sedimentary Accumulation, from the Standpoint of Historical Geology)," *Izv. A. N. S.S.S.R.*, ser. Geol., no. 2, 1946.

449. Strakhov, N. M., "O periodichnosti i neobratimoy evolyutsii osadkobrazovaniya v istorii zemli (On the Periodicity and the Progressive Evolution of the Deposition of Sediments in the Earth's History)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 6, 1949.
450. Strakhov, N. M., *Osnovy istoricheskoy geologii (Foundations of Historical Geology)*, Gosgeolizdat, 1948, vols. I-II.
451. Strakhov, N. M., "Zakonomernosti orogeneza v osveshchenii Shtille (The Laws of Orogenesis, as Explained by Stille)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 10, no. 3-4, 1932.
452. Studer, B., *Lehrbuch der physikalischen Geographie und Geologie*, bd. II, 1847.
453. Suess, E., *Das Antlitz der Erde*, bd. I-III, 1883-1909.
454. Suess, E., *Die Entstehung der Alpen*, 1875.
455. T'an, H., "A Summary of the Geological History of Szechuan and Sikard," *Bull. Soc. Geol. China*, vol. 16, 1936-1937.
456. Teichmueller, R., and J. Schneider, "Die Grenze von Alpen und Apennin," *Abh. Ges. Wiss. zu Goettingen, Math-phys. Kl.*, folge III, h. 14, 1935.
457. "Tektonika tsentral'nogo kazakhstana (Tectonics of Central Kazakhstan)," chap. 1 in *Tektonika S.S.S.R.*, vol. 1, A. N. S.S.S.R., 1948.
458. Tetyayev, M. M., *Osnovy geotektoniki (Foundations of Geotectonics)*, 1934.
459. Tetyayev, M. M., *Osnovy Geotektoniki (Foundations of Geotectonics)*, 2d ed., 1941.
460. Tikhomirov, V. V., "Malyy kavkaz v verkhnemelovoye vremya (osnovnyye tipe otlozheniy i usloviya ikh obrazovaniya) [The Lesser Caucasus in Late Cretaceous Times (The Basic Types of Deposits and the Conditions of Their Formation)]," *Tr. In-ta Geol. Nauk A. N. S.S.S.R.*, ser. Geol., vyp. 123, no. 44, 1950.
461. Tikhomirov, V. V., "K voprosy o razvitii zemnoy kory i prirode granita (The Problem of the Development of the Earth's Crust and the Nature of Granite)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 8, 1958.
462. Tikhomirov, V. V., and V. Ye. Khain, *Kratkiy ocherk istorii geologii (Brief Survey of Geologic History)*, 1956.
463. Ting, V., "The Orogenic Movements in China," *Bull. Geol. Soc. China*, vol. 8, no. 2, 1929.
464. Tolstoy, J., and W. M. Ewing, "North Atlantic Hydrography and the Mid-Atlantic Ridge," *G. S. A. Bull.*, vol. 60, no. 5, 1949.
465. Tolstoy, J., "Submarine Topography in the North Atlantic," *G. S. A. Bull.*, vol. 62, no. 5, 1951.
466. *Trudy pervogo soveshchaniya po voprosam kosmogonii 16-19 aprelya 1951 g (Publications of the First Conference on Problems of Cosmogony, 16-19 April, 1951)*, A. N. S.S.S.R., 1951.
467. Tsuboi, Chuji, "Deformations of the Earth's Crust as Disclosed by Geodetic Measurements," in *Ergebnisse der kosmischen Physik*, vol. IV, 1939.
468. Turner, F. J., *Mineralogical and Structural Evolution of the Metamorphic Rocks*, G. S. A. Memoir 30, 1948.

469. Twenhofel, U. H., *Ucheniye ob obrazovanii osadkov (The Theory of the Formation of Sediments)*, ONTI, 1936.
470. Ul'yanov, A. V., "Moshchnosti svit kak kriteriy pri opredelenii perspektiv neftenosnosti (The Thicknesses of Sedimentary Strata as a Criterion of the Prospects for Finding Oil)," *Sov. Geol.*, vol. 9, 1946.
471. Umbgrove, J. H. F., *The Pulse of the Earth*, 2d ed., 1947.
472. Urey, H. C., "On the Origin of Continents and Mountains," *Proc. Nat. Acad. Sci.*, vol. 39, pp. 933-946, 1953.
473. Usov, M. A., "Elementy tektoniki leninskogo rayona kuznetskogo kamennougol'nogo basseyna (Elements of the Tectonics of the Lenin District in the Kuznets Coal Basin)," *Izv. Sib. Otd. Geol. Kom.*, t. 5, vyp. 4, 1926.
474. Usov, M. A., "Sostav i tektonika kemerovskogo mestorozhdeniya kuznetskogo kamennougol'nogo basseyna (The Composition and Tectonics of the Kemerovskoye Deposit in the Kuznets Coal Basin)," *Izv. Sib. Otd. Geol. Kom.*, t. 5, vyp. 4, 1926.
475. Usov, M. A., "Fatsii i formatsii izverzhennykh gornyykh porod (Facies and Formations of Extrusive Rocks)," in *Voprosy geologii sibirii (Problems of the Geology of Siberia)*, A. N. S.S.S.R., 1945.
476. Usov, M. A., *Fazy i tsikly tektogeneza zapadno-sibirskogo kraya (Phases and Cycles of Tectogenesis in the Western Siberian Region)*, 1936.
477. Usov, M. A., "Fazy tektogeneza (The Phases of Tectogenesis)," *Sov. Geol.*, no. 11, 1938.
478. Usov, M. A., "Formy diz'yunktivnykh dislokatsiy v rudnikakh kuzbassa (The Types of Disjunctive Dislocations in the Ore Bodies of the Kuznets Basin)," in *Sbornik po Geol. Sibiri.*, Tomsk, 1933.
479. Usov, M. A., *Strukturnaya geologiya (Structural Geology)*, Gosgeolizdat, 1940.
480. Usov, M. A., "Geotektonicheskaya teoriya samorazvitiya materii zemli (The Geotectonic Theory of the Development of the Material in the Earth)," *Izv. A. N. S.S.S.R.*, ser. Geol., vyp. 1, 1940.
481. Usov, M. A., "Tektonika anzherskogo kamennougol'nogo mestorozhdeniya (The Tectonics of the Anzherskoye Coal Deposit)," *Izv. Sib. Geol. Kom.*, vol. 1, vyp. 4, 1920.
482. Usov, M. A., Tektonika sudzhenskogo kamennougol'nogo mestorozhdeniya (The Tectonics of the Sudzhenskoye Coal Deposit), *Izv. Sib. Geol. Kom.*, vol. 1, vyp. 2, 1919.
483. Uzhik, G. V., *Soprotivleniye otryzu i prochnost' metallov (The Ultimate Strength of Metals and Their Resistance to Breaking)*, Izd. A. N. S.S.S.R., 1950.
484. van Bemmelen, R. W., *The Geology of Indonesia*, vol. I, The Hague, 1949.
485. van Bemmelen, R. W., *Mountain Building*, The Hague, 1954.
486. van Bemmelen, R. W., "Tectogenèse par gravité," *Bull. Soc. Belge de géologie*, t. 69, f. 1, 1955.
487. van Bemmelen, R. W., "The Undation Theory of Development of the Earth's Crust," *Trans. Int. Geol. Congr.*, Washington, vol. 2, 1933

488. van Bemmelen, R., "The Undation Theory . . . ," *Rep. XVI Sess. Int. Geol. Cong., U.S.A., 1933*, vol. 2, 1936.
489. van der Gracht, W. A. J. M., "Die jungpalaeozoische Gebirgsbildung in den Alpen," *Vierteljahrsschrift Naturf. Ges. Zürich*, vol. 64, 1919.
490. van der Gracht, W. A. J. M., "The Structure of the Salt Domes of Northwest Europe as Revealed in Salt Mines," *Bull. A. A. P. G.*, no. 2, 1925.
491. Varentsov, M. I., *Geologicheskoye stroyeniye zapadnoy chasti kurinskoy depressii (The Geologic Structure of the Western Part of the Kura Depression)*, Izd. A. N. S.S.S.R., 1950.
492. Vassoyevich, N. B., "K voprosu ob usloviyakh obrazovaniya flisha (On the Conditions Governing the Formation of a Flysch)," *Izv. A. N. S.S.S.R., ser. Geol.*, vyp. 4, 1940.
493. Vassoyevich, N. B., *Flish i metodika yego izucheniya (Flysch and the Methods of Its Study)*, Lengostoptekhzdat, 1946.
494. Vassoyevich, N. B., *Usloviya obrazovaniya flisha (Conditions of Flysch Formation)*, ONTI, 1951.
495. Veber, V. V., "Geologicheskaya karta kabristana: Planshet II-3 (bayan-ata) [Geologic Map of Kabristan: Sheet II-3 (Bayan-Ata)]," *Tr. NGRI*, vyp. 62, 1935.
496. Vening-Meinesz, F. A., "Convection Currents in the Earth and the Origin of the Continents," *Proc. Koninkl. Nederl. Akad. Wet.*, vol. 55, no. 5, ser. B, 1952.
497. Vening-Meinesz, F. A., "Earth's Crust Deformation in Geosynclines," *Proc. Koninkl. Nederl. Akad. Wet.*, vol. 53, no. 1, 1950.
498. Vereshchagin, G. Yu., "K voprosu o neravnomernosti podnyatiya beregov onezhskogo ozera (On the Nonuniform Uplift of the Shores of Lake Onega)," *Gos. Gidrol. In-t. Tr. Olonetskoy Nauchn. Eksped.*, ch. III, vyp. 2, 1931.
499. Vereshchagin, G. Yu., "Polozhitel'nye i otritsatelnyye dvizheniya beregovoy linii na oz. segozero (Advances and Retreats of the Shorelines of Lake Segozero)," *Gos. Gidrol. In-t. Tr. Olonetskoy Nauchn. Eksped.*, ch. III, vyp. 1, 1926.
500. Voynovskiy-Kruger, K. G., "O shirine ugleobrazuyushchey zony (The Width of the Coal-forming Zone)," *Sov. Geol.*, vol. 38, 1949.
501. Wadia, D., *Geology of India*, 1926.
502. Wegener, A., *Proiskhozhdeniye materikov i okeanov (The Origin of the Continents and the Oceans)*, GIZ, 1924.
503. Wegmann, C. E., "Zur Deutung der Migmatite," *Geol. Rundschau*, bd. 26, 1935.
504. Wettstein, A., "Über die Fischfauna des Tertiaerer Glarnerschiefers," *Abh. Schweiz. psl. Ges.*, vol. 13, 1886.
505. Williams, M. J., *Arisaig-Antigonish District, Nova Scotia*, Geol. Surv. Canada Memoir 60, 1914.
506. Willis, B., *The Mechanics of Appalachian Structures: 13th Annual Report*, U.S.G.S., 1893.
507. Willis, B., and R. Willis, *Strukturnaya Geologiya (Structural Geology)*, translated from the 2d American edition, Nefteizdat, Baku, 1932.

508. Willis, B., and R. Willis, "Eruptivity and Mountain Building," *G. S. A. Bull.*, vol. 52, no. 10, 1941.
509. Wong, W., "Crustal Movements and Igneous Activity in Eastern China since Mesozoic Time," *Bull. Geol. Soc. China*, vol. 6, no. 1, 1927.
510. Wong, W., "The Mesozoic Orogenic Movements in Eastern China," *Bull. Geol. Soc. China*, vol. 7, no. 1, 1929.
511. Yanshin, A. L., "O pogrebennykh gertsinidakh k vostoku ot kaspiyskogo morya (On the Buried Hercynian Structures in the Region to the East of the Caspian Sea)," *Byull. Mosk. Ob-va Isp. Pr., Otd. Geol.*, vol. 21, no. 5-6, 1945.
512. Yeliseyev, N. A., et al., "Geologo-petrograficheskiy ocherk khibinskikh tundr (Geologic and Petrographic Survey of the Khibinsk Tundras)," in *Mezhd. Geol. Kongress, XVII Sess., severnaya ekskursiya, kol'skiy poluostrov*, 1937.
513. Zakharov, Ye. Ye., and N. I. Korolev, *Struktura rudnogo polya, mineralogicheskiiy sostav i genezis nikitovskogo rtutnogo mestorozhdeniya v donetskom bassejne (The Structure of the Ore District, the Mineral Composition and the Genesis of the Nikitovka Mercury-ore Deposit in the Donets Basin)*, A. N. S.S.S.R., 1940.
514. Zavaritskiy, A. N., "Nekotoryye fakty, kotoryye nado uchityvat' pri tektonicheskikh postroyeniyyakh (Some Facts That Must Be Taken into Account in Tectonic Constructs)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 2, 1946.
515. Zavaritskiy, A. N., *Vvedeniye v petrokхимию (Introduction to Petrochemistry)*, Gosgeolizdat, 1944.
516. Zenkovich, V. P., *Dinamika i morfologiya morskikh beregov: I. Volnovyye protsessy (The Dynamics and Morphology of Seashores: I. Wave Processes)*, Morskoy Transport, 1946.
517. Zenkovich, V. P., "K voprosu o vyrabotke rel'yefa podvodnoy platformy (The Reworking of the Relief of a Submarine Platform)," *Izv. A. N. S.S.S.R., ser. Geogr. i Geof., vyp. 2*, 1940.
518. Zhemchuzhnikov, Yu. A., "Sloy i plast (Bed and Stratum)," *Izv. A. N. S.S.S.R., ser. Geol.*, no. 5, 1950.
519. Zittel, K. A., *Geschichte der Geologie und Paläontologie*, 1899.

INDEX OF GEOGRAPHIC NAMES

NOTE: The bank or tributary of a river is designated as right or left as it would appear facing downstream.

Abkhaziya—an Autonomous Republic at the Black Sea end of the Caucasus Mountains.

Abul-Samsarskaya Ridge—a ridge trending north-south in the Lesser Caucasus, long. 43°20'E.

Akhalkalaki—a city, long. 44°20'E, lat. 41°55'N.

Akhtsu Range—a range in the Caucasus Mountains, near Sochi.

Aksubayevo—a village, long. 50°40'E, lat. 55°10'N.

Alakul, Lake—a lake, long. 82°E, lat. 46°N.

- Alayskaya Valley—a valley between the Alayskiy and Zaalayskiy mountain ranges.
- Alayskiy Range—one of the southern spurs of the Tien Shan.
- Altay Mountains—mountains at the boundaries between the U.S.S.R., China, and Mongolia.
- Assa River—a river on the northern slopes of the Greater Caucasus, a right tributary of the Sunzha River, in the Terek River basin.
- Balkhash, Lake—a lake in the southeastern part of the Kazakh folded region, long. 74–79°E, lat. 45–47°N.
- Belaya River—a river in the northern Caucasus, a left tributary of the Kuban River.
- Bol'shoy Balkan—mountains on the eastern shore of the Caspian Sea.
- Bugulma—a city, long. 53°E, lat. 54°30'N.
- Chatkal'skiy Range—one of the western spurs of the Tien Shan.
- Chegem River—a right tributary of the Baksan River, in the Terek River basin.
- Cheshskaya Bay—a gulf of the Barents Sea.
- Chingiz (Chingiz-Tau) Range—a mountain range in eastern Kazakhstan, north of Lake Balkhash.
- Chulym River—a right tributary of the Ob' River in western Siberia.
- Chuya River—a right tributary of the Katun River, on the upper reaches of the Ob' River basin.
- Dagestan—an Autonomous Republic in the northeastern Caucasus.
- Darvakh Range—a mountain range of the Pamir system.
- Dzhezkazgan—a city in Kazakhstan, long. 67°E, lat. 48°10'N.
- Fisht, Mt.—a mountain in the Caucasus, long. 39°55'E, lat. 43°55'N.
- Gagry—a city on the Caucasus shore of the Black Sea, long. 40°15'E, lat. 43°20'N.
- Gissarskiy Range—a mountain range in the eastern Tien Shan.
- Gornaya Shoriya—a mountainous area in the northern Altay.
- Irtys River—a left tributary of the Ob' River in western Siberia.
- Ishimbay—a city on the Belaya River, long. 56°E, lat. 53°27'N.
- Issyk-Kul', Lake—a lake in the central Tien Shan, south of Alma-Ata.
- Izh River—a right tributary of the Kama River in the Volga River basin, in the southern Urals.
- Kabristan—a region southeast of the Caucasus, adjoining the Apsheron Peninsula.
- Kalba Range—a mountain range in Kazakhstan, not far from the Altay.
- Karabakh—an upland area southeast of the Lesser Caucasus.
- Karaganda—a city in Kazakhstan, long. 73°30'E, lat. 50°N.
- Kara-Kum—a desert in central Asia at the borders between the U.S.S.R., Irak, and Afghanistan.
- Kara-Tau Range—a mountain range in the southern Urals, east of the city of Ufa.
- Kara-Tau Range—one of the northwestern spurs of the Tien Shan.
- Karateginskiy Range—a western spur of the Pamir Mountains.
- Kerch—a peninsula between the Black Sea and the Sea of Azov.
- Khatanga River—a river in eastern Siberia, emptying into the Laptev Sea.

Kirgiz Range—the northern range of the Tien Shan.

Kirovobad—a city in the Transcaucasus (in the Azerbaydzhan S.S.R.).

Kirovograd—a city, long. $32^{\circ}30'E$, lat. $47^{\circ}N$.

Klyaz'ma River—a left tributary of the Oka River, in the Volga River basin.

Kolkhida—a region at the Caucasus end of the Black Sea, around the mouth of the Rion River.

Krasnaya Polyana—a settlement in the Caucasus, long. $40^{\circ}12'E$, lat. $43^{\circ}40'N$.

Kungey Alatau Range—a mountain range in the Tien Shan system, along the northern shore of Lake Issyk-Kul'.

Kuraminskiye Mountains—the southwestern range of the Tien Shan.

Kusarskiy Plain—an alluvial plain in the southeastern Caucasus.

Kuznets Basin—a coal basin in the southeastern part of western Siberia.

Kuznetskiy Alatau Range—the northern range of the Altay mountain system.

Kyzyl-Kum—a sand desert in central Asia, southeast of the Aral Sea.

Malaya (and Bol'shaya) Kinel' River—right tributaries of the Samara River in the Volga River basin.

Malka River—a left tributary of the Terek River in the northern Caucasus.

Mangyshlak Peninsula—a peninsula on the eastern shore of the Caspian Sea.

Maykop—a city in the northern Caucasus, long. $40^{\circ}07'E$, lat. $44^{\circ}35'N$.

Maykublen—a mine in central Kazakhstan.

Medveditsa River—a left tributary of the Don River.

Minusinsk—a city on the Yenisey River, long. $91^{\circ}40'E$, lat. $53^{\circ}43'N$.

Mugodzhary Mountains—a southern spur of the Urals.

Muya River—a left tributary of the Vitim River, in the Lena River basin in eastern Siberia.

Naberezhnyye Chelny—a village near the eastern shore of the Caspian Sea, long. $52^{\circ}30'E$, lat. $55^{\circ}40'N$.

Naryn Depression—a depression on the upper reaches of the Naryn River in the Tien Shan.

Novo-Lyalinsk (Novaya Lyalya)—a city in the central Urals, long. $60^{\circ}40'E$, lat. $59^{\circ}05'N$.

Nuratinskiy (Nura-Tau) Range—one of the western spurs of the Tien Shan.

Peter I Range—a mountain range in the Pamir system.

Pshekha River—a left tributary of the Belaya River, in the Kuban River basin of the western Caucasus.

Pskemskiy Range—one of the northwestern spurs of the Tien Shan.

Pur River—a river in western Siberia, emptying into the Ob' River estuary.

Rudnyy Altay—a region in the western foothills of the Altay mountain system.

Salair Range (Salairskiy Ridge)—a range of mountains southeast of the city of Novosibirsk.

Segozero, Lake—a lake in northwestern Russia, long. $33^{\circ}10'-34^{\circ}15'E$, lat. $63^{\circ}10'-63^{\circ}30'N$.

Serov—a city in the central Urals, long. $60^{\circ}35'E$, lat. $59^{\circ}35'N$.

Sevan, Lake—a lake in the Lesser Caucasus, long. $45^{\circ}00'-45^{\circ}40'E$, lat. $40^{\circ}10'-40^{\circ}35'N$.

Shakh-Dag—a mountain in the southeastern part of the Greater Caucasus, long. $48^{\circ}E$, lat. $41^{\circ}20'N$.

Sheshma River—a left tributary of the Kama River in the Volga River basin.

- Sok River—a left tributary of the Volga River, north of the city of Kuybyshev.
- Stepnoy Zay River—a left tributary of the Kama River, in the Volga River basin.
- Sundyr' River—a right tributary of the Volga River, near the city of Cheboksara.
- Sunzha Range—one of the northern spurs of the eastern Caucasus.
- Tadzhik Depression—a depression in the Pyanzh River valley, along the border between the U.S.S.R. and Afghanistan, in central Asia.
- Tagil'sk—a city in the northern Urals.
- Talasskiy Alatau Range—the northern mountain range of the Tien Shan system.
- Talyshskiy Range—a range of mountains trending northwest, along the boundary between the U.S.S.R. and Iran.
- Taz River—a river in western Siberia, emptying into the Ob' River estuary.
- Tengiz, Lake—a lake in Kazakhstan, long. 69°E, lat. 50°30'N.
- Tera Range—one of the northern spurs of the Caucasus.
- Terek River—a river that runs down the northern slopes of the Caucasus and empties into the Caspian Sea.
- Timan or Tinanskiy Ridge—a ridge of mountains in the Northern European part of the U.S.S.R.
- Tom' River—a right tributary of the Ob' River, at the city of Tomsk.
- Trialetskiye Mountains—a range in the Lesser Caucasus system.
- Tsna River—a right tributary of the Oka River, in the Volga River basin.
- Tuarkyr—an upland region east of the Gulf of Kara-Bogaz-Gol on the Caspian Sea.
- Tunguska Basin—a coal basin in the western part of the Siberian platform.
- Turan Lowland—a lowland area east of the Aral Sea.
- Turgay (Turgayskaya) Valley—a valley trending north-south between the low hills of the Urals and Kazakhstan, long. 64°50'E.
- Turkestan Range—one of the western spurs of the southern Tien Shan.
- Ugamskiy Range—one of the western spurs of the Tien Shan.
- Ukhta River—a left tributary of the Izhma River in the Pechora River basin, in the Northern European part of the U.S.S.R.
- Ulutau—a mountain range in the western part of central Kazakhstan.
- Ust'-Urt—a plateau between the Caspian and Aral Seas.
- Vilyuy River—a left tributary of the Lena River in eastern Siberia.
- Vostochnyye Sayany Range—a mountain range trending northwest, west of Lake Baykal in the southern part of Siberia.
- Vyatka River—a left tributary of the Volga River.
- Yelabuga—a city west of the Urals, long. 52°10'E, lat. 55°50'N.
- Zaalaitskiy Range—the northern mountain range in the Pamir system.
- Zailiyskiy Alatau Range—the northern mountain range in the western Tien Shan system, north of Lake Issyk-Kul'.
- Zapadnyy Sayany Range—a mountain range trending northeast, west of Lake Baykal in the southern part of Siberia.
- Zeravshan Range—a mountain range in the western part of the southern Tien Shan.
- Zidda River—a river in the Pamir Mountains.
- Zilair—a city in the southern Urals, long. 58°E, lat. 52°N.

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